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HIGHLAND-LOWLAND INTERACTIONS IN THE GANGES BRAHMAPUTRA RIVER BASIN: A REVIEW OF PUBLISHED LITERATURE

L.A. Bruijnzeel with C.N. Bremmer

ICIMOD OCCASIONAL PAPER NO. 11

Kathmandu, Nepal

May, 1989

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ACKNOWLEDGEMENTS

The present report is the result of a chain of events: early 1987, I received a request from the "Ganges Working Group", based in Utrecht, the Netherlands, for information on the relationships between land-use practices in the Himalaya and flooding and siltation in the Indo-Gangetic plain. This request was prompted by the concern of these people about the so-called Dutch Inland Water Transport Project, which aimed at improving the navigability of the Ganges between Allahabad and Haldia. The discussion centered around the question to what extent reforestation of the uplands would have any effect on siltation and flooding in the lowlands, and therefore on the feasibility of the said project.

Although I had collected a fair amount of information on the Himalayan environment over the years, I was not aware of any in-depth study of "highland-lowland interactions", based on quantitative information rather than opinions. Christiaan Bremmer, a graduate student in my department, undertook the task of sorting out the literature and writing a very readable account, entitled: "The role of vegetation and land use in flooding, erosion and mass wasting in the Ganges drainage basin, India" as part of his studies in hydrology.

I ventured to send a copy of the review to ICIMOD for comments in the fall of 1987. Then came "the" flood in Bangladesh, and with it the invitation from ICIMOD to convert Bremmer's draft review into an ICIMOD Occasional Paper. Upon arrival in Kathmandu in late December 1988, I felt the need to expand the geographical scope of the study and so the Tsangpo-Brahmaputra basin was included as well. This of course meant that a great deal of new literature had to be included and much of January 1988 was spent in libraries in Kathmandu. Since then, a completely new report has been written, mainly in Amsterdam, based on about three times as many literature references as the original review contained.

It is a great pleasure to record the help received by a large number of individuals, both in Nepal and in the Netherlands. Without their support this report would not have been possible.

First of all I wish to thank Dr. Colin Rosser, Director of ICIMOD, for the opportunity to write this paper and visit Nepal on two occasions, and for his support throughout the course of the undertaking. The weeks spent at ICIMOD were very happy ones indeed, not in the least because of the interaction with its staff, notably Dr. Don Alford, Dr. Jayanta Bandyopadhyay, Professor Suresh Chalise, Dr. Anis Dani and Dr. Kk. Panday, all of whom readily shared their time, knowledge and sense of humour with me on many occasions. Professor Li Tian-chi, head of the Division of Mountain Environmental Management, and his administrative staff were most helpful in getting countless pieces of information xeroxed. It will be hard to forget your smiles. The ingenuity displayed by Mr. R.B. Shrestha (ICIMOD library) in obtaining some of the documents on which this report is based, proved invaluable. I would also like to thank Mr. Surendra Shrestha and Mrs. Priya Trosuwan for their smooth logistic arrangements.

Brian Carson, Vic Galay, Don Gilmour, Rob Gerritsen, Wolfgang Grabs, P. van Groen, G.J. Klaassen, Peter Laban, Jan Rupke and P.B. Shah all provided literature, slides or other information, which is gratefully acknowledged.

Romée de Vries is thanked for typing the bulk of the report, Saskia Kars and R.Joshi for their skillful draughtswork, Maarten Waterloo for computerized diagrams, and Mrs. Prabha Thacker and Professor Ian Simmers for editorial comments.

Last and certainly not least I wish to express my gratitude to that lady whose support over the past few months has been beyond a simple thank you, and to Alexander M. Smit for helping to keep the right perspective.

Sampurno Bruijnzeel

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I INTRODUCTION

The combined basins of the Ganges and Brahmaputra rivers in the northern part of the Indian sub-continent (1.38 million square kilometres) constitute the home of some 400 million people, or about 10% of the world's population. By modern standards most of the population of the region has always been poor, but concern about the diminishing welfare of these people, as a result of environmental degradation, has been growing rapidly during the last decade (Eckholm, 1975; Hrabovsky & Miyan, 1987).

A widely held perception of this environmental degradation involves recent massive deforestation in the uplands as a result of a population explosion, leading in turn to catastrophic increases in soil erosion and river sediment loads on the one hand and to greatly increased flooding and siltation in the lowlands, on the other (Bowonder, 1982; Spears, 1982; Myers, 1986). This view is reflected in the frequently quoted statement that a few tens of a million Himalayan hill farmers are holding hostage several hundred million inhabitants of the plains (World Resources Institute, 1985). Naturally, the economic and political implications of the above scenario are enormous (Ives, 1987).

Despite the fact that this view appeals to logic and conventional wisdom alike, it has been challenged in the past few years (e.g. Carson, 1985; Hamilton, 1987). Basically this alternative school of thought stresses that the dramatic geophysical processes, responsible for the very existence of the Himalaya and the plains, are sufficiently impressive and that the effects of the activities of mountain farmers are insignificant in comparison. In this view, whereas the mountain environment is indeed being harmed through loss of forest, - creating grave local problems, this does not appreciably increase flood danger or sedimentation in the densely populated plains. The image of the people living in the lowlands as "hostages" would thus be highly overdrawn.

As pointed out by Ives (1987), the controversy is largely a matter of scale and historical perspective. A crisis of very large dimensions *is* developing, but the real magnitude of the problem has been disguised by oversimplifications and generalizations. The recent Mohonk Mountain conference (Ives & Ives, 1987) on "the Himalaya-Ganges problem" was intended to expose this simplistic approach, but concluded at the same time that there is an urgent need for a much fuller understanding of Himalaya-Ganges dynamics (Ives, 1987).

The main purpose of the present paper is to shed more light on the above debate by reviewing the evidence with respect to the physical aspects of the matter published to date. The basic question, therefore, is: "What is the role of forest and land-use in the uplands with respect to flooding, dry-season flows and sedimentation in the lowlands?". And, following immediately: "What downstream benefits can be reasonably expected in this regard from upland reforestation?".

By drawing as much as possible on quantitative data actually collected in the region itself, rather than extrapolating results obtained in other (usually less extreme) parts of the world, we hope to lift the discussion from the level of empiricism and subjectivity to a more objective presentation of scientifically established facts. Naturally, the success or failure of such an approach depends to a large extent on the amount and reliability of the information collected.

Before the issue at hand can be addressed in more detail, one needs to have an idea of the spatial and temporal variations of the relevant environmental elements, notably geology and geomorphology, climate, soils and vegetation, as well as their interaction. Chapter II, therefore, describes the physical and biological characteristics of the environment in some detail. Since water is the key element linking atmosphere, plants and soils, this is followed by a discus-

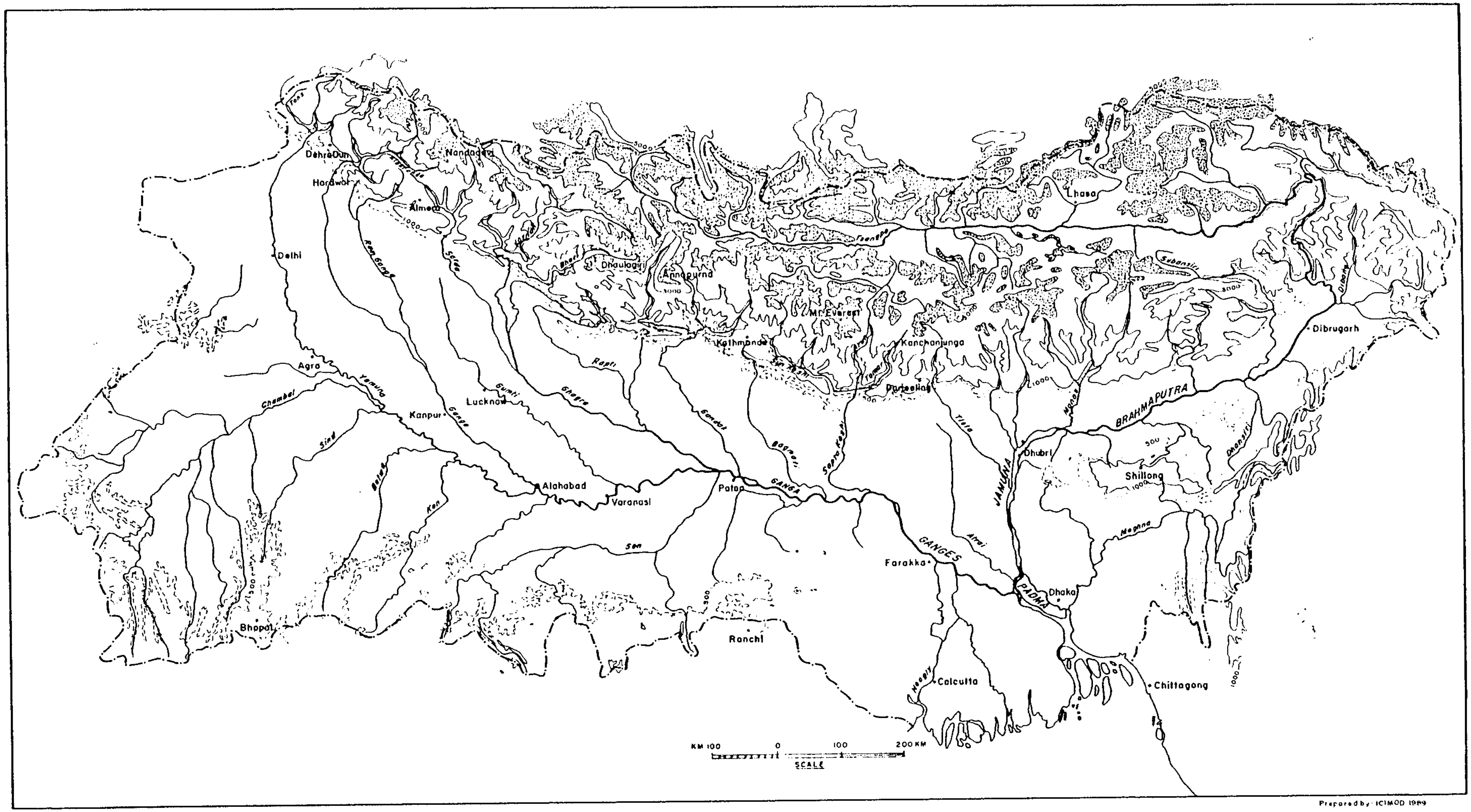


Figure 1 The Ganges-Brahmaputra River Basin: general features.

sion of regional hydrology (Chapter III). Within the framework thus provided, the specific influences exerted by various land-use types on total and seasonal water yields, surface erosion, mass movement processes and river sediment loads, are dealt with at length in Chapter IV. Based on these findings, conclusions are drawn and gaps in our knowledge indicated (Chapter V).

II MOUNTAINS AND PLAINS

II.1 THE MAIN RIVER SYSTEMS

Although the two major rivers after which the basin is named have their origins within 250 km from each other on either side of the Himalaya, and their outlets adjacent to one another in the northernmost part of the Bay of Bengal, their courses show important contrasts (Figures 1 and 2).

The *Ganges* rises south of the main Himalayan divide near Gangotri in the Indian state of Uttar Pradesh. Initially it runs in southerly directions and at a steep gradient (Figure 2a), before breaking through the outer Himalayas near Dehradun and reaching the WNW-ESE oriented plain that is named after it at Hardwar, some 350 km from the source.

From here onwards, the Ganges continues to flow at ever smaller gradients, viz. from about 100 cm/km in the piedmont zone around Hardwar to less than 6 cm/km in Bihar (Figure 2a). Meandering as well as braiding occurs here (Singh & Verma, 1987).

Between Hardwar and the river's mouth (a stretch of 2600 km) the Ganges is joined by numerous tributaries, both from the young mountain range in the north as well as from the old plateau in the south (Figure 1). The most important tributary is the Yamuna, which drains more than one third of the entire Ganges basin, including a significant portion of the southwestern plateau. It joins the Ganges at Allahabad. The major Himalayan tributaries are the Karnali (draining 12% of the basin) and the Sapt Kosi (9%). The latter river is particularly notorious for its unreliable character and claims have been made that this river alone contributes 40% of the entire sediment load of the Ganges (Alford, 1988a). The most important tributary from the south is the Son, which covers almost 7% of the total area, has a fairly steep gradient (45 cm/km on average) and a rather flashy flow regime (Singh & Singh, 1987). Near Farakka the Ganges splits up into several branches, the most important of these being the Hooghly (which reaches the Bay of

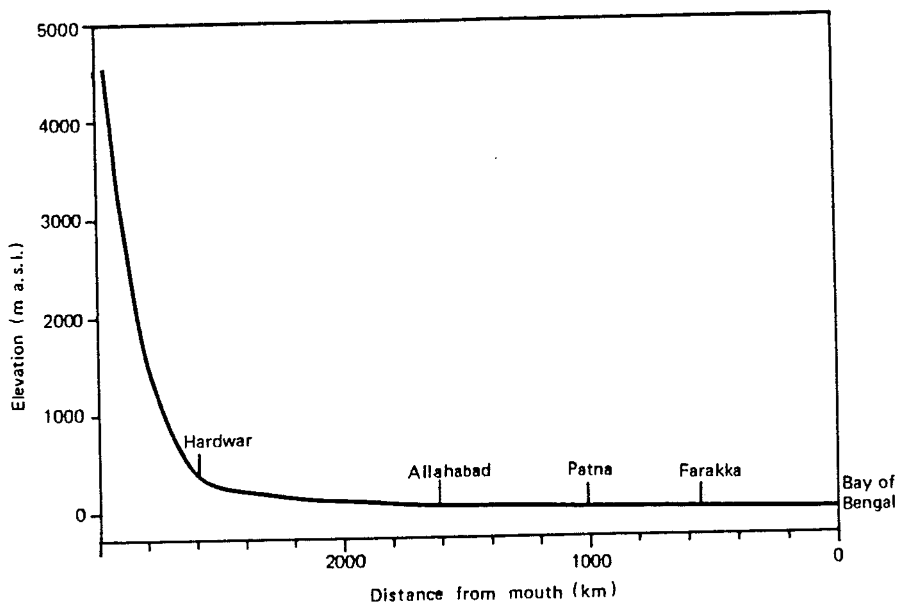


Figure 2a **Approximate longitudinal profile of the river Ganges as determined from small-scale maps.**

Bengal at Calcutta), the Madhumati, and the Ganges proper. The latter eventually joins the Brahmaputra and is then called Padma (see below).

The ***Brahmaputra*** (or Tsangpo as it is called in Tibet) on the other hand, originates in the Kailas range north of the high Himalaya. It does not break through the mountains immediately, but flows eastward for about 1200 km and at a fairly low gradient (Figure 2b) along the bottom of a flat tectonic valley parallel to and about 160 km north of the Himalaya (Figure 1). At the extreme eastern end of its course in Tibet, the Tsangpo suddenly takes a turn to the south and cuts a deep and narrow gorge upon crossing the high Himalayan range. The gradient of the river in the gorge section is extremely steep (Figure 2b). The river then traverses another 225 km of mountainous terrain before debouching onto the Assam plain near Pasighat at an elevation of only 155 m. After being joined by the Dibang and Lohit rivers, the combined flow, which is now called Brahmaputra, moves westward through the valley of Assam for about 720 km, becoming strongly braided in response to the greatly reduced slope of its bed (Figure 2b; Goswami, 1985). Before turning south again at Dhubri, the Brahmaputra receives major contributions from a number of Himalayan and other rivers (Figure 1).

Shortly after crossing the border with Bangladesh, the river is joined by the Tista, which, like the Kosi, for centuries has been notorious for its capacity to flood. The Brahmaputra then bifurcates: the smaller (eastern) channel retains the name Brahmaputra and the main channel is now called Jamuna. Some 200 km further south the Jamuna joins forces with the Ganges, the combined flow now being called Padma. The latter (and also the

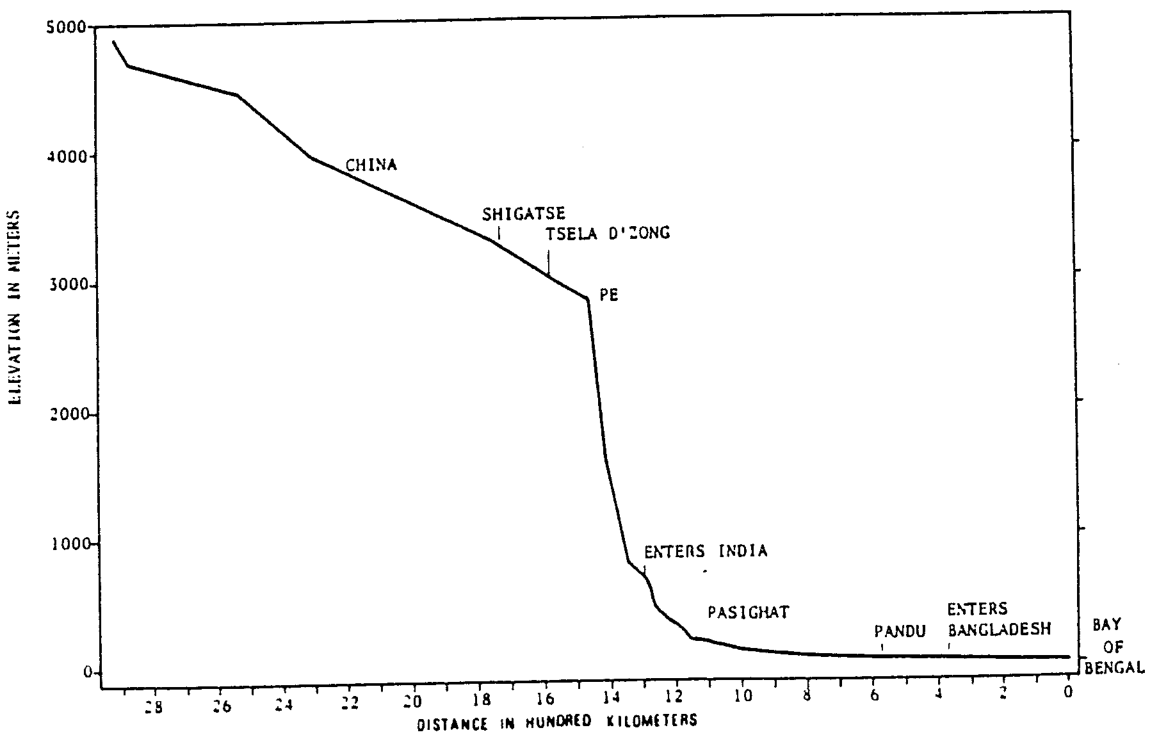


Figure 2b **Longitudinal bank profile of the Brahmaputra River (after Goswami, 1985).**

Brahmaputra) eventually becomes part of the Meghna river system, which drains most of eastern Bangladesh and the surrounding hills, before discharging their waters into the Bay of Bengal (Figure 1).

II.2 GEOLOGY AND GEOMORPHOLOGY

II.2.1 Ganges River Basin

A broad distinction can be made between the old and geologically stable shield area in the south, the Indo-Gangetic depression in the centre, and the young and geologically active Himalayan mountain range in the north (Figure 3).

Southern Plateau

The southern plateau consists largely of very old (Palaeozoic and older) crystalline and sedimentary rocks, in some parts overlain by younger (Mesozoic) sedimentary rocks, and, especially in the southwestern part of the basin (Figure 5), by the hard lava flows known as the Deccan traps. Although the shield occasionally attains maximum heights of about 1000 m (Figure 1), most of the area lies between 300 and 600 m. This not only reflects the mature character of the region, but also moderate sub-recent geological uplift associated with the Himalayan orogeny. As such, rivers emanating from the area have somewhat higher gradients than would be expected for such an old landscape (Saxena, 1987). In addition, these rivers have flashy flow regimes and have shifted their beds upon entering the Ganges plain for thousands of years (Singh & Singh, 1987). However, despite very serious surface erosion in the area (Samuel & Das, 1982), sediment loads carried by most southern tributaries of the Ganges are only a fraction of those carried by the northern tributaries (Gupta, 1975; see section III.6).

Himalaya

As for the Himalayan range, a frequently used sub-division, which is more or less applicable over the entire length of the chain as covered by the present study, follows that proposed for Nepal by Hagen (1969) (Figures 3 and 4).
The various units are roughly parallel to each other and from north to south consist of:

(1) the Tibetan Marginal Range and Plateau, sometimes called Tethys- or Trans-Himalaya (and known as Darma-Johar in the western parts of the mountains);
(2) the Great or High Himalaya (called Himadri in the west);
(3) the Middle or Lesser Himalaya, and (4) the Siwalik foothills.

The Middle Himalaya consists of the Fore-Himalaya (also known as the High Mountains), the central Midlands, and the Mahabharat Range (Figure 4). The latter two are sometimes referred to as "Middle Hills" (Nepal) or "Himachal" (India).

The boundaries between the four major units are mainly of a geological nature. Two important features in this respect are the so-called "main boundary thrust" (MBT) and the "main central thrust" (MCT) (Figure 4).

The general features of the Himalaya can perhaps be explained best in the context of plate tectonics (Stöcklin, 1980). The plate of the Indian sub-continent passes underneath the main Eurasian plate, thereby elevating the Tibetan plateau. At the same time the Himalayan mountain range is created, consisting essentially of the scrapings from the forward edge of the Indian plate forced back (i.e. to the south) over the advancing mass (hence the northern dips of the MBT and MCT).

Rates of uplift are in the order of 1 mm/yr (Zeitler et al., 1982; Iwata et al., 1984). "Slippage" in the process of uplift regularly generates tremors and earthquakes.

In the following pages the four major units will be described briefly (Gansser, 1964; Rupke, 1974; Stöcklin, 1980; Carson et al., 1986; Figures 3-5).

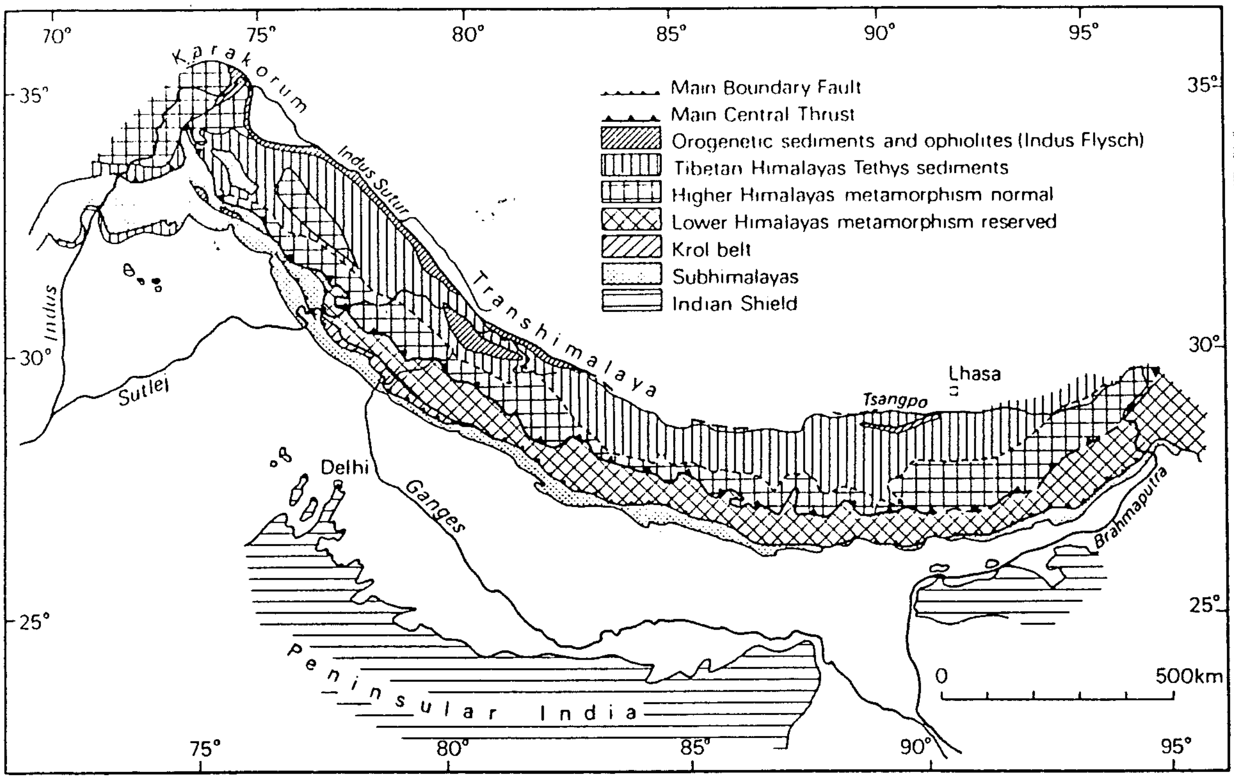


Figure 3 Main geo-structural features of the Ganges-Brahmaputra River Basin (after Gansser, 1964).

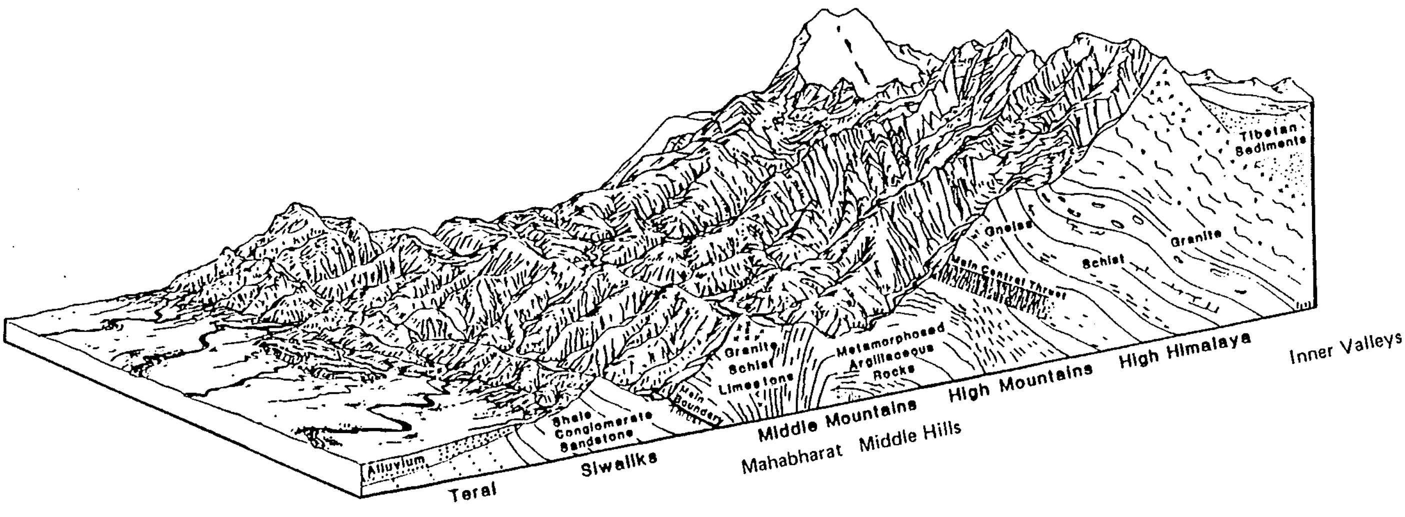


Figure 4 Physiographic zones of the Nepal Himalaya (Galay, 1987; modified from Nelson et al., 1980, and Ramsay, 1986)

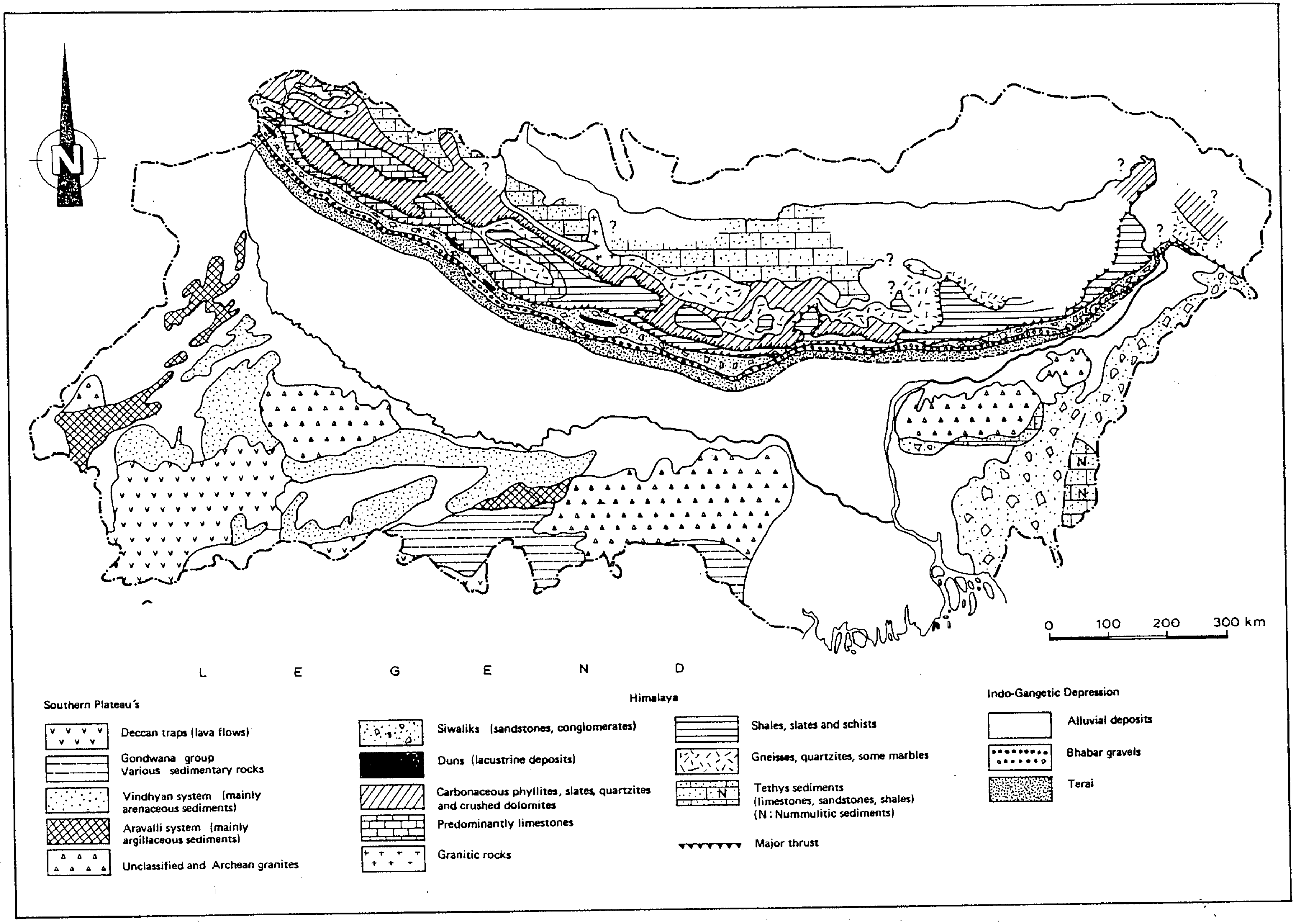


Figure 5. Generalized geological map of the Ganges-Brahmaputra River Basin (modified from Gansser, 1964; Fuchs, 1980).

The *Tibetan Marginal Range* consists largely of sedimentary rocks, such as limestones and shales, that have been little affected by metamorphism. Topography is strongly controlled by the lithology of the underlying rocks. Situated in the rain-shadow of the great peaks, annual precipitation totals are low (see II.3.1) and vegetation, where present, is sparse and xerophytic. As such, gullying, initiated and maintained by occasional showers, is widespread in areas with soft deposits (Plate 13).

In view of the low precipitation and high elevation, physical weathering processes (rock fall/frost shattering, solifluction or debris flows on slopes wetted by snowmelt, undercutting by rivers, etc.) predominate over chemical weathering, resulting in poorly developed and shallow soils. A number of the most important tributaries of the Ganges (e.g. Karnali, Gandaki, Arun) originate in this zone, breaking through the entire mountain chain. Such rivers are called antecedent as they were in existence before the Himalayan orogeny. As such, these river systems are millions of years old. They generally follow Palaeozoic fault systems that have been preserved throughout the orogeny (Plate 1).

Plate 1 **The upper Gandaki river near Kagbeni, southern Mustang, Nepal. Note the various levels of river terraces corresponding with various phases of uplift and river incision.**

The *Great Himalaya* physiographic region includes all the great peaks and consists predominantly of metamorphic rocks. The most resistant of these (gneisses, quartzites, siliceous marbles, occasionally granites and certain types of schists) often constitute the highest parts. The entire region is subject to intense physical weathering. In addition to most of the processes indicated for the Tibetan Marginal Range, the High Himalaya is characterized by active glaciation. Glaciers, cirque bases, U-shaped valleys, hanging tributary valleys, moraines and avalanche slopes are all common landforms. More than 80% of the region has bedrock at or near the surface on very steep terrain (Plate 2).

Plate 2 **Physical weathering processes are extremely active in the glaciated environment of the High Himalaya (North face of Annapurna, Manang district, Nepal).**

The next unit, the *Fore Himalaya* (or High Mountains) is essentially distinguished from the Great Himalaya by the presence of a smaller amount of snowpack, reflecting lower elevations because of slightly less resistant (but also metamorphic) rocks. All valleys in this region have been glaciated in the past. Active river down-cutting since the glaciation has produced impressive canyons and high relief is common. River gradients are high to very high (cf. Figure 2b). Mass movements are frequent, especially along waterways and on the steeper slopes (Plate 3). Moderate gully erosion sometimes results from snowmelt, whereas sheet erosion is induced occasionally by overgrazing of alpine grasslands on moderate to steep slopes.

Soils are coarse-textured and often shallow, reflecting the cool climate and resistant bedrock. Trees, when present, are generally coniferous and often scattered and give way to grassland at higher elevations (3500-4000 m).

Plate 3 **Upon crossing the High Himalaya, rivers such as the Bhote Kosi in Central Nepal, have carved deep gorges. Note the active erosion and mass-movement processes in the riparian zone.**

The Main Central Thrust (Figure 4), although a fairly irregular zone rather than a sharp line (Figure 5), forms a well-defined boundary between the relatively high Fore Himalaya and the much lower (600-2000 m) *Central Midlands*. As the MCTZ is highly fractured, it is also the scene of intense mass movement processes (see Chapter IV). The Central Midlands generally consist of weakly to moderately metamorphosed rocks that are less resistant to erosion than the more highly metamorphosed rocks exposed further north (Plate 4). Consequently, slopes may be gentler, especially on the relatively soft phyllites. Since the region has not experienced significant glaciation and because of the warmer climatic conditions prevailing here, fairly deep (often reddish) soils have developed in the phyllites.

Steeper slopes (and shallower soils) are associated with harder rock types, such as quartzites and limestones, which are especially common in the western part of this zone (Figure 5). The lowest parts of the Central Midlands are found in structurally controlled valleys, which often contain extensive ancient lake or river terraces (Plate 5). The latter are remnants of past river damming due to huge landslides, or may represent the deposits of large glacial outburst floods before the last glaciation (Carson 1985).

Plate 4

Typical view of the Middle Hill region of Central Nepal near Dhulikhel. Note gully erosion and frequent landslide scars in grazing land.

Plate 5

Land use and erosion processes in the Kathmandu valley, Middle Hill region, Central Nepal. Note frequent shallow landslips in scrubland and old major slide in river bend in background.

Gradients of the main rivers flowing through the region are much gentler than in the High Himalaya and many tributaries form extensive alluvial fans near confluences. River bank cutting is also a widespread phenomenon in these valleys. Surface erosion is minimal, even on the steepest slopes, as long as they are covered with the original sub-tropical and warm-temperate forests. However, considerable parts of this relatively densely populated region are now occupied by agricultural fields and more or less severely degraded scrub and grasslands (see Section II.4; Plate 4). The consequences for hillslope hydrological response to rainfall and intensity of erosion processes associated with such changes in land-use, will be discussed in Chapter IV.

In contrast to surface erosion, mass movements have always been common on steep (> 30 °) slopes in this region (Plate 4). There are indications, however, that the number of shallow (less than 3 m deep) slides is increasing following vegetation removal, even on less steep slopes (Carson, 1985; Euphrat, 1987; Manandhar & Khanal, 1988; see Chapter IV).

The third and southernmost unit of the Middle Himalaya is called ***Mahabharat Lekh*** in Nepal (Figure 4). It rises to heights of upto 2500 m and is made up of rock types that are more competent (i.e. resistant to erosion) than those of the Central Midlands (although they are much less metamorphosed than the hard rocks of the High Himalaya). This leads to steep topography and often relatively thin soils. Common lithologies are quartzites, schists, gneisses, and (especially in the western parts) limestones and dolomites (Figure 5). Deeper soils can be found on river terraces and on fractured granite (Carson et al., 1986).

The Mahabharat range is geologically the most complex of the various zones, consisting of a number of thrust sheets ("nappes") pushed towards the south. When these were created, they presented a barrier to the rivers flowing south from the High Himalaya and many rivers changed their courses in a westerly or easterly direction (Figure 1) before finding a weak zone and breaking through. River gradients in the Mahabharat Lekh are again higher than in the Central Midlands.

Due to the often very steep topography and the presence of only a shallow veneer of colluvium over resistant bedrock, natural erosion by mass movement processes is high (Carson et al., 1986; Plate 6).

The range also constitutes a barrier to moist air masses coming from the south and slopes with northern aspects receive much less rainfall than slopes with southern aspects (Figure 10). Therefore surface erosion hazards are variable. Forest cover (broad-leaved and coniferous) in this zone is generally much better than in the Central Hills.

South of the Main Boundary Fault the Middle Himalaya abruptly gives way to a series of relatively low (up to 1200 m) ridges called the <u>Siwaliks</u> (Figure 4). These consist of Tertiary unconsolidated and highly erodible fluviatile sediments ranging from relatively fine-grained graywackes in the south ("Lower Siwaliks"), through soft sandstones interbedded with thin layers of clay ("Middle Siwaliks"), to very coarse sands and conglomerates ("Upper Siwaliks") in the north (Figure 6a).

The maximum thickness of the Siwalik formation has been estimated at over 4000 m in Nepal (Weidner, 1981) to about 5500 m in the Kumaon Himalaya (Rupke, 1974). As such they bear testimony to the long standing erosion history of the Himalayan range. Although relative differences in elevation throughout the Siwaliks are generally much less than 1000 m, the landscape is quite rugged and its river system exceedingly flashy (Singh et al., 1982).

In the Nepalese Siwaliks, about ten times more area was mapped as dry river channels than in the Middle Himalaya (Carson et al., 1986; Plate 7), despite the fact that much of the area is still forested (mostly

Plate 6

Deep-seated slides on well-forested hillslopes in the Mahabharat range of Central Nepal. Slides of this size are capable of temporarily damming a river, which upon bursting may produce a devastating surge of water and sediment.

Plate 7

Although the Siwaliks in general still have good forest cover, pressures on this very vulnerable ecosystem are increasing. Note the remarkably wide valley bottoms with braided rivers (Central Nepal)

Shorea, but more scrubby in the drier parts of the region). Burning and grazing of the understorey is widely practised. In combination with the weakly structured nature of the soils and the prevailing high rainfall intensities, this has led to intense surface erosion, e.g. in the vicinity of Chandigarh (Pant, 1983). In addition, natural slope instability is such, that some of the area has been rated as the most unstable in all of Nepal (Carson et al., 1986).

An important feature of the Siwalik range is the occurrence of rather broad E-W oriented valleys (*"Duns"*), filled with lacustrine sediments deposited at a time when rivers were obstructed in their flow by the rise of the Siwaliks (Plate 8). Erosion from the hillsides has produced numerous alluvial deposits, which, like the lacustrine sediments, present more or less severe surface erosion and gullying hazards, depending on cover and slope steepness (Nelson et al., 1980). The Siwalik zone is separated from the Indo-Gangetic depression in the south by an active thrust fault (Figure 6a).

Indo-Gangetic depression

The Indian plate has been bent downward by the weight of the Himalaya, thus creating a major structural basin, filled with debris from the mountains in the north, and to a lesser extent from the southern plateau. The "northern" deposits are often grouped into a piedmont (footslope) zone and more lowlying alluvial deposits. The entire complex is commonly referred to as the ***Terai***, although the upper part of the piedmont is known as the ***Bhabar*** zone (100-300 m a.s.l., Figure 5).

The Bhabar formation essentially consists of very coarse-textured alluvial fan deposits topped by a thin layer of finer textured material. The thickness of the sediment is at a maximum close to the Siwaliks and may reach 5000 m (Carson et al., 1986). Soils are well to excessively drained and pose drought problems. Forest clearance has been limited up to now, mainly for this reason, but where it has occurred, serious sheet and gully erosion has been the result. Because of the dramatic reduction in river gradients upon leaving the Siwaliks, streams, which are heavily laden with sediment during the monsoon, assume a braided pattern (Plate 9). River beds can be up to several kilometres wide and erosion is lateral rather than vertical (cf. Section III.6). Indeed, the piedmont cone south of the Himalaya is one of the most impressive (and active!) of any such system on earth (Plate 9).

Much of the water infiltrating into the coarse Bhabar deposits emerges a few kilometres downstream, producing a sudden change in river morphology from braided to meandering (Figure 6a). This coincides with a knickpoint in river gradient, with the deposits becoming increasingly finer textured to the south (Weidner, 1981). Also, drainage density increases considerably below this "line of saturation". From here onwards (in the Terai proper), a system of sandy levees and more clayey depressions ("basins") is found, which bears no relationship to the present-day drainage pattern (Figures 6b & c), although the depressions get flooded occasionally. Rice is widely cultivated in these "basins", whereas the (higher) levees are used for settlements and roads.

In between the old basin/levee system and the active river channels, another sub-recent system exists, called the Meander Floodplain (Figure 6c). Its channels are only flooded during extreme events (Weidner, 1981).

In contrast to all other physiographic zones described above, the Terai is a zone of deposition rather than erosion. In addition, it suffers greatly from the rapid and unpredictable shifting of river beds (Carson, 1985; Section III.6). The sub-tropical forests of the Terai are being converted to agricultural fields at a rapid rate (Section II.4.2).

The alluvial deposits of the ***Ganges plain*** are mainly composed of unconsolidated beds of sand and gravel (former channel beds), silt and clay

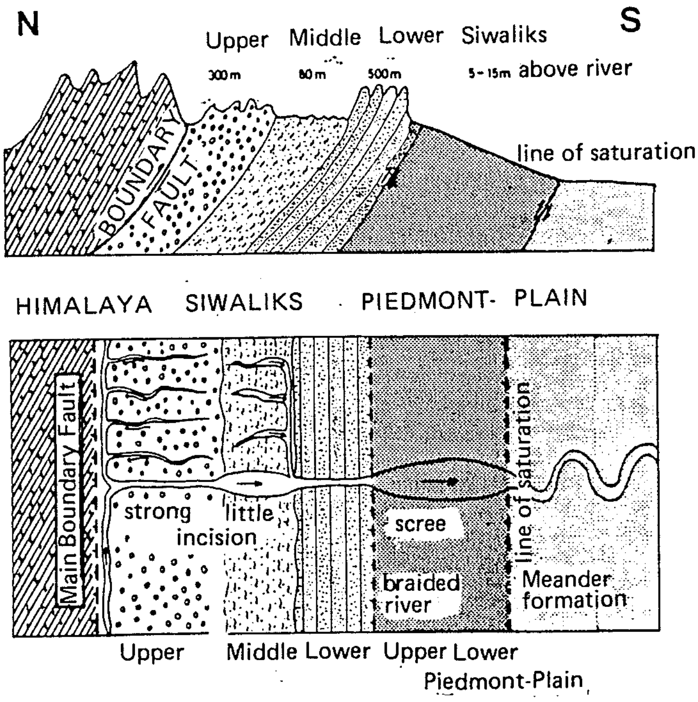
Plate 8

A Dun valley in Central Nepal. To the right the well-forested steeply sloping beds of the Upper Siwaliks, to the left the intensively cultivated southern margin of the Mahabharat. Note the huge landslide scars in the foreground.

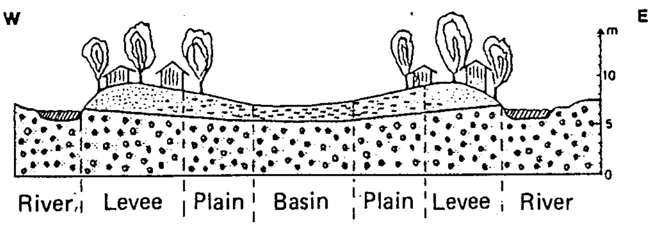
Plate 9

Upon entering the piedmont (Bhabar) zone, the Himalayan rivers deposit large amounts of coarse material and continue to flow through unstable channels.

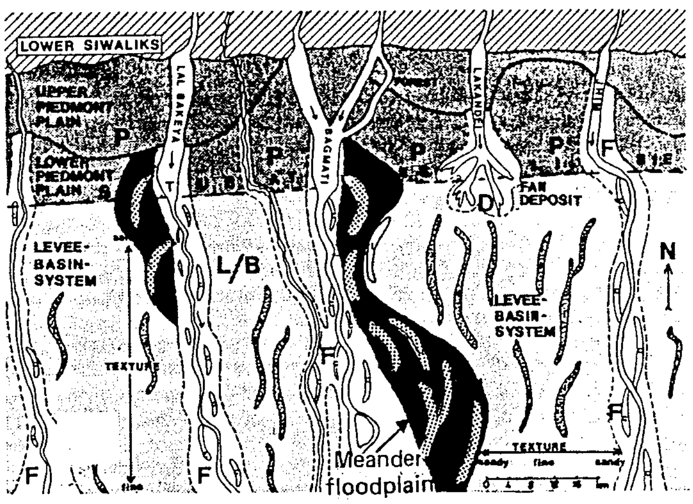
Figure 6. (after Weidner, 1981).



(a). Geology, geomorphology and river dynamics of the Siwalik-Piedmont zone.



(b). Cross section of the Levee-Basin system of the Terai .



(c). Spatial distribution of relief units distinguished in 6a and b.

(former depressions), and their mixture in varying proportions. Although the alluvial fitting on average is 1300-1400 m deep, decreasing gradually southward, a zone of over 8000 m (!) depth runs along the foot of the Himalayas (Singh & Verma, 1987).

A distinction is made between the *Bhangar*, or high (> 15 m above the plains) interfluvial zones above the general limit of flooding, and the *Khadar*, the more lowlying riverine tracts whose sandy to clayey deposits are annually renewed.

II.2.2 Brahmaputra River Basin

Excluding the deltaic parts for the moment, the Brahmaputra River Basin can be subdivided into four major physiographic units, viz. the old Meghalaya tableland and the younger Patkai-Naga (or Purvanchal) ranges in the south, the central Assam valley depression, and the eastern extension of the Himalaya in the north (Figure 3).

The *Meghalaya plateau* rises to elevations of 600 m in the west to about 2000 m in the centre and consists mainly of very old (Pre-Cambrian) hard crystalline rocks (granites and the like). Along the western and southern margins flat-lying sedimentary rocks (dominated by sand- and limestones of Mesozoic to Tertiary age) are exposed. Whereas the central and eastern parts form a true plateau, the western and northern fringes are highly dissected, forming a series of irregular low hills down to the Assam valley (Das et al., 1987).

To the east of the plateau, and separated from it and the Assam valley by a major fault, the *Purvanchal* is found. It consists of a strongly fractured series of N-S to NE-SW trending ridges. Rocks comprise a variety of Tertiary sediments, topography is steep and river valleys narrow. In addition, the area is extremely unstable in terms of seismicity due to its proximity to the main transform fault along which the Indian and South-East Asian tectonic plates rub shoulders (Haroun er Rashid, 1977). Elevations range from a low 150 m in the southwest to over 3000 m in the extreme northeast. Most of the area, however, is found between 900 and 2100 m (Singh & Mukherjee, 1987).

The *Assam valley* is an almost flat plain underlain by some 1500 m of alluvium. Its width ranges from about 90 km at the upstream end to about 60 km lower down. Within the plain a number of isolated granitic hillocks, that have become detached from the Meghalaya plateau, are found. As can be expected from the contrasts in geology between the mountains in the north and in the south, the physiography of the two river banks differs markedly, especially in the western part of the valley.

In the north, a situation similar to that described earlier for the Terai exists, with braided streams that start to meander upon passing the line of saturation. However, before joining the Brahmaputra, these rivers run almost parallel to the main stream as they encounter its levees. In the south on the other hand, the valley is much less wide and the small tributaries flowing from the Meghalaya plateau run in much less meandering courses (Figure 1).

The *Eastern Himalaya* is runs through Sikkim, Bhutan and Arunachal Pradesh. The general division of Siwaliks (locally called *Duars*), Middle and Great Himalaya and Tibetan Marginal Range still applies here, although the topography of the Middle Himalayas now rises steadily and merges with the Great Himalaya. As such the latter do not stand out as much as they do in Nepal and further west (Jangpangi, 1978). Also the depression associated with the Central Midlands is hardly developed here.

Instead, there is a growing tendency for spurs from the Great Himalaya to radiate southward (e.g. the Black Mountains in Bhutan) as one moves towards the east. This is in line with the general change of direction in the axis of the mountains as they approach the eastern end of the Indian tectonic plate.

The *Bengal Basin* (elaborately described by Haroun er Rashid, 1977) has been filled with sediments washed down from the surrounding highlands, mostly since Pleistocene times. Morphologically speaking, the area consists of the floodplains of the rivers traversing it, with all the features common to such systems (Plate 10), ending up in a major delta (Figure 1).

Although the basin as a whole is a zone of deposition, it is nevertheless (like the Indo-Gangetic depression) subject to tectonic movements, with some parts actively sinking and others rising. Indeed, the change in the course of the Brahmaputra in 1787 during a single flood event has been described as having become possible because of such movements (Morgan & McIntire, 1959). Estimates of the rates of sinking vary from 2 to 6 cm/yr (Haroun er Rashid, 1977). Needless to say that such areas have become even more liable to flooding than before (Section III.5).

II.3 CLIMATE

II.3.1 Spatial and seasonal variations in precipitation

The climate of the Indian sub-continent and the Himalaya is dominated by the monsoon, the seasonal reversal of winds and associated rain. During the sun's annual march, the differential heating of the various latitudinal zones generates air movements, whose general annual patterns are quite predictable (Figure 7). However, as we shall see later, there are significant and unpredictable deviations from the overall pattern (Mooley & Parthasarathy, 1983).

In the Indian Ocean region the dominant pattern is one of high-level winds that blow NE to SW between October and June. During this time relatively little precipitation is generated.

Plate 10 View of the Jamuna floodplain between Dacca and Bahadurabad, Bangladesh (photograph by G.J. Klaassen).

Figure 7a.

Normal dates of the onset of the southwest monsoon (after Rao, 1981).

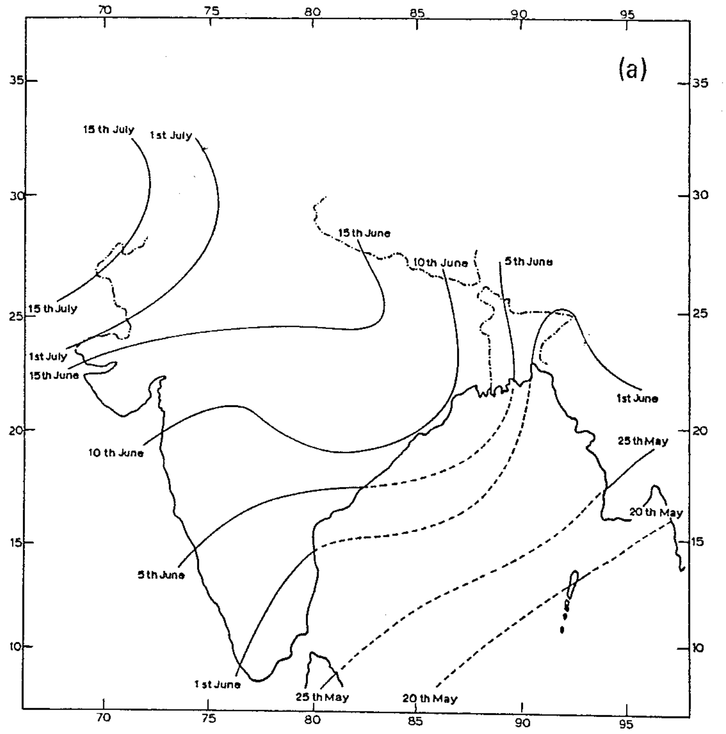
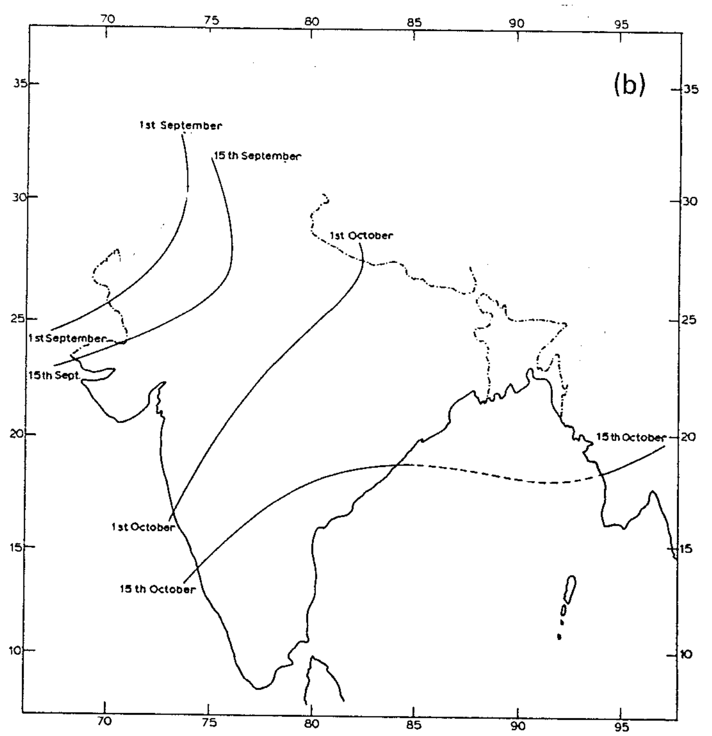


Figure 7b.

Normal dates of the withdrawal of the southwest monsoon (after Rao, 1981).



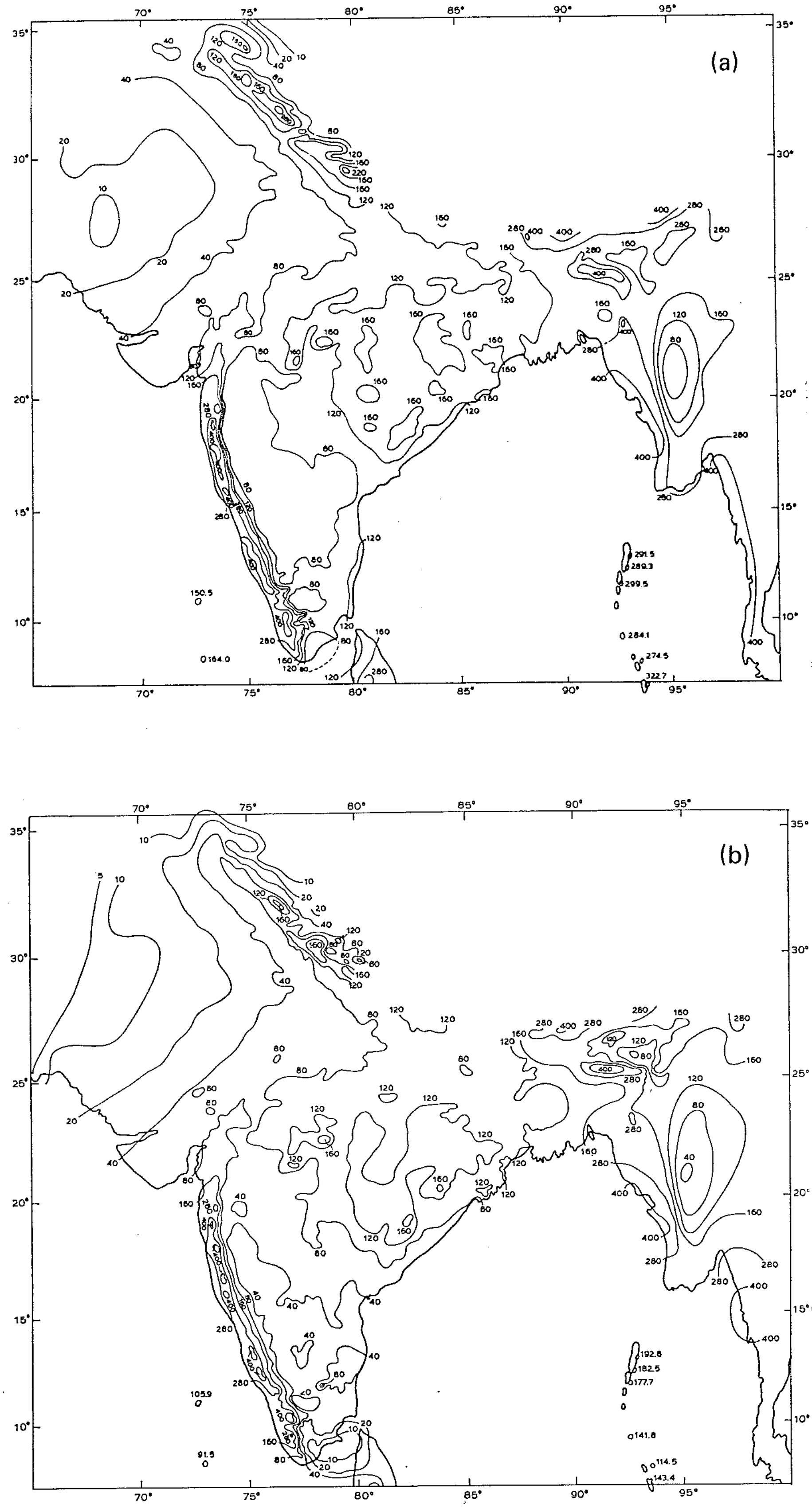


Figure 8. **Mean annual (a) and summer monsoon (b) rainfall totals (cm) over the Indian sub-continent (after Rao, 1981).**

This changes to a predominant movement of upper air from SW to NE about three months after the sun has begun its march into the northern hemisphere. Coming from the Indian Ocean, the southwesterly winds are warm and laden with moisture. As the air rises upon reaching the land mass, it cools off, whereby its moisture-holding capacity is reduced. Heavy rainfall is the result all over the sub-continent between June and September.

In fact, rainfall over much of the area is concentrated so much during these months, that the isohyetal patterns for the monsoon and the annual totals are very similar indeed (Figure 8).

A glance at Figures 7 and 8 reveals that not only is the eastern part of the combined river basin subjected to the influence of the monsoon for a longer time, it also receives considerably higher rainfall totals than most of the western parts. Also, the number of raindays (with rain > 2.5 mm) per year is significantly higher in the east (ranging from ca. 75 in Bangladesh to 100-150 in Assam), as compared to the central (ca. 50) and western (50-75) portions of the basin (Rao, 1981).

Apart from this overall trend of lesser rainfall towards the west, there are marked topographical effects producing strong local variations throughout the entire river basin.

For example, some of the world's highest rainfall totals have been recorded at Cherrapunji on the southern slopes of the Meghalaya plateau, where the moist air from the Bay of Bengal suddenly rises by 1200 m (Rao, 1981). At Shillong, situated only 50 km to the north on the same plateau, rainfall has already decreased to about 2400 mm/yr, declining steadily to 1600-1800 mm/yr as one descends into the Assam valley. It rises again to 4000 mm/yr on the lower Himalayan slopes of Arunachal Pradesh (Figure 9a).

Little is known about the behaviour of rainfall in the higher parts of this region (Rao, 1981; Goswami, 1985), but recent data from Bhutan (Figure 9b; Sharma, 1985) suggest a strong decline in precipitation with elevation. As shown below, this pattern deviates somewhat from that observed further west.

The various longitudinal physiographic zones of the Himalaya, with their strong contrasts in elevation (Figure 4), experience widely varying amounts of rainfall (Figures 10 and 11). To start with (taking Nepal as an example), there is an increase in annual rainfall from the southern Terai (ca. 1500 mm) to the Siwalik range (over 2000 mm). North of the Siwalik and Mahabharat mountains, there are several areas where annual precipitation falls below 1500 or even 1000 mm due to their sheltered positions which deprives them of moisture-laden southerly winds. Rainfall then increases again along the lower slopes of the Great Himalaya (Figure 10).

Maximum totals are experienced around Pokhara, at the foot of the Annapurna Himal, where topography rises dramatically over a very short distance, and to a lesser extent in the upper Arun basin between the Khumbu and Kanchenjunga Himals (Figure 11).

Regionally therefore, elevation may strongly influence rainfall totals, sometimes even to the extent that the general east-west trend is superseded (Figure 11). These spatially varying amounts of rainfall represent an even stronger variation in rainfall erosive power, or erosivity (Chapter IV.2).

Statistically significant relationships between annual rainfall and elevation have been derived for some parts of the Lesser Himalayas (Meyerink, 1974; Kathyar & Striffler, 1984; Ramsay, 1985), although some investigators have been unable to detect any such pattern (e.g. Dittmann, 1970).

It should be realised that such relationships have only local predictive value. In addition, they only hold for a certain range in altitude. Above a critical elevation (2500-3500 m: Upadhyay & Bahadur, 1982; Dobremez, 1976), a reduction in

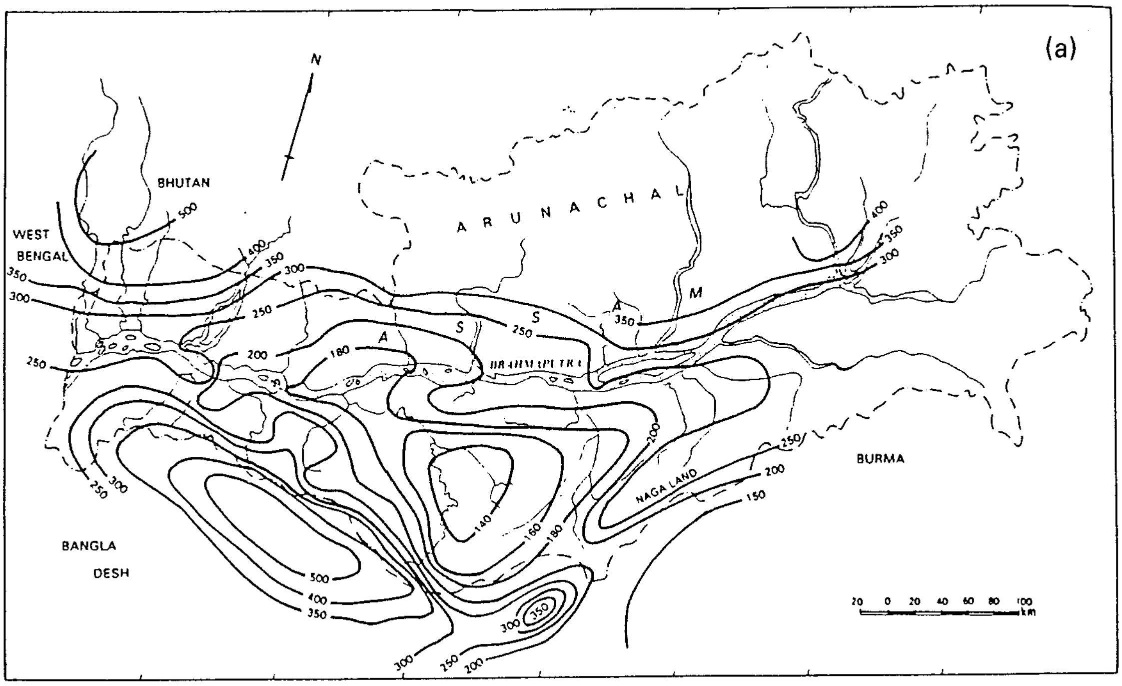


Figure 9a Isohyetal map of the Assam valley and adjoining highlands (after Goswami, 1985).

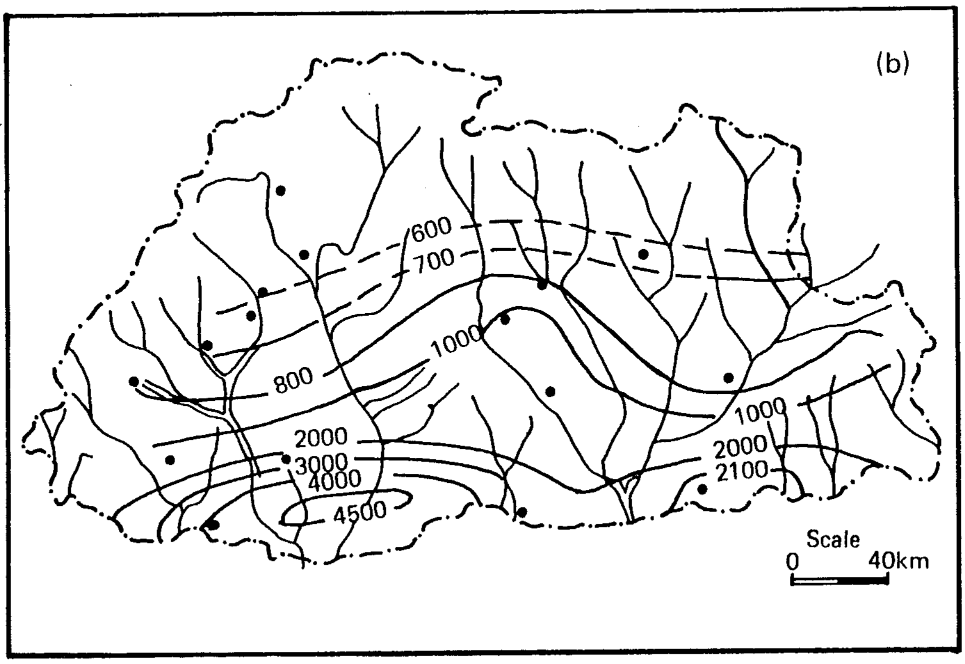


Figure 9b Isohyetal map of Bhutan (modified from Sharma, 1985).

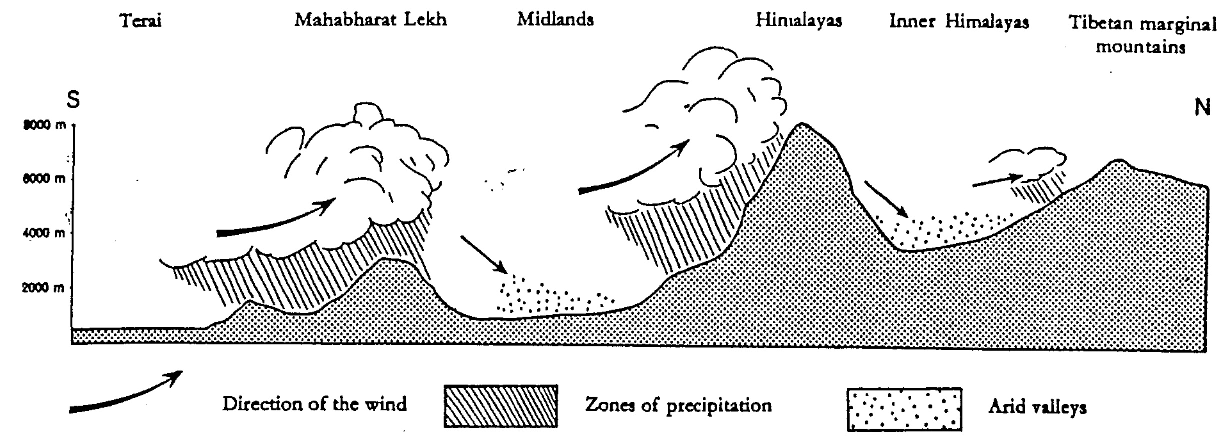


Figure 10 Principal zones of precipitation in the Nepalese Himalaya (after Hagen, 1969).

precipitation is often observed (cf. Figure 11).

North of the Great Himalaya, precipitation totals also decrease rapidly to less than 500 mm/yr in the Dolpa and Mustang districts.

Equally low amounts of precipitation have been reported for the Tibetan plateau (e.g. Lhasa: 360 mm/yr; Majupuria & Majupuria, 1988).

A very similar overall pattern was observed by Dhar et al. (1987) for the Kumaon and Garhwal Himalayas, although there precipitation totals rarely exceeded 2000 mm/yr.

Superimposed on these regional to sub-regional variations in precipitation in the mountains, there are important differences on a local scale, for example between valley bottoms and exposed ridges (Domroes, 1979; Higuchi et al., 1982). Very little work has been carried out in this respect, but results collected to date suggest (much) lower precipitation totals for valley bottoms (where the majority of the rainfall stations are located) as compared to slopes and ridges (Higuchi et al., 1982). The dry-valley bottom effect has been ascribed to the occurrence of persistently ascending mountain breezes (Flohn, 1970).

This observation not only has important agricultural and hydrological implications, but also suggests that regional precipitation totals as presented on isohyetal maps, etc. often will be underestimates.

A minimum rain-gauge network density of one gauge per 100 km^2 has been suggested for areas experiencing convectional rainfall and orographic effects (Chyurlia, 1984). Such densities, however much desirable, are nowhere near approximated anywhere in the Himalaya.

Taking Nepal again as an example, the total number of rainfall stations was 138 (1/1070 km^2) in 1975, the majority of which (83%) was located below 2000 m a.s.l. (Dobremez, 1976). However, considerable progress has been made since then, with the total number of stations in 1984 already amounting to ca. 300 (1/490 km^2), fifty of these (17%) located above 2000 m and thirty above 2500 m a.s.l. (HMG Ministry of Water Resources, 1986).

It will be clear from the above, that information on the amounts of precipitation falling as *snow* is particularly limited. Chyurlia (1984) reported a weak increase in snowfall with elevation for three stations in Nepal. In addition, he estimated seasonal variations in snowline

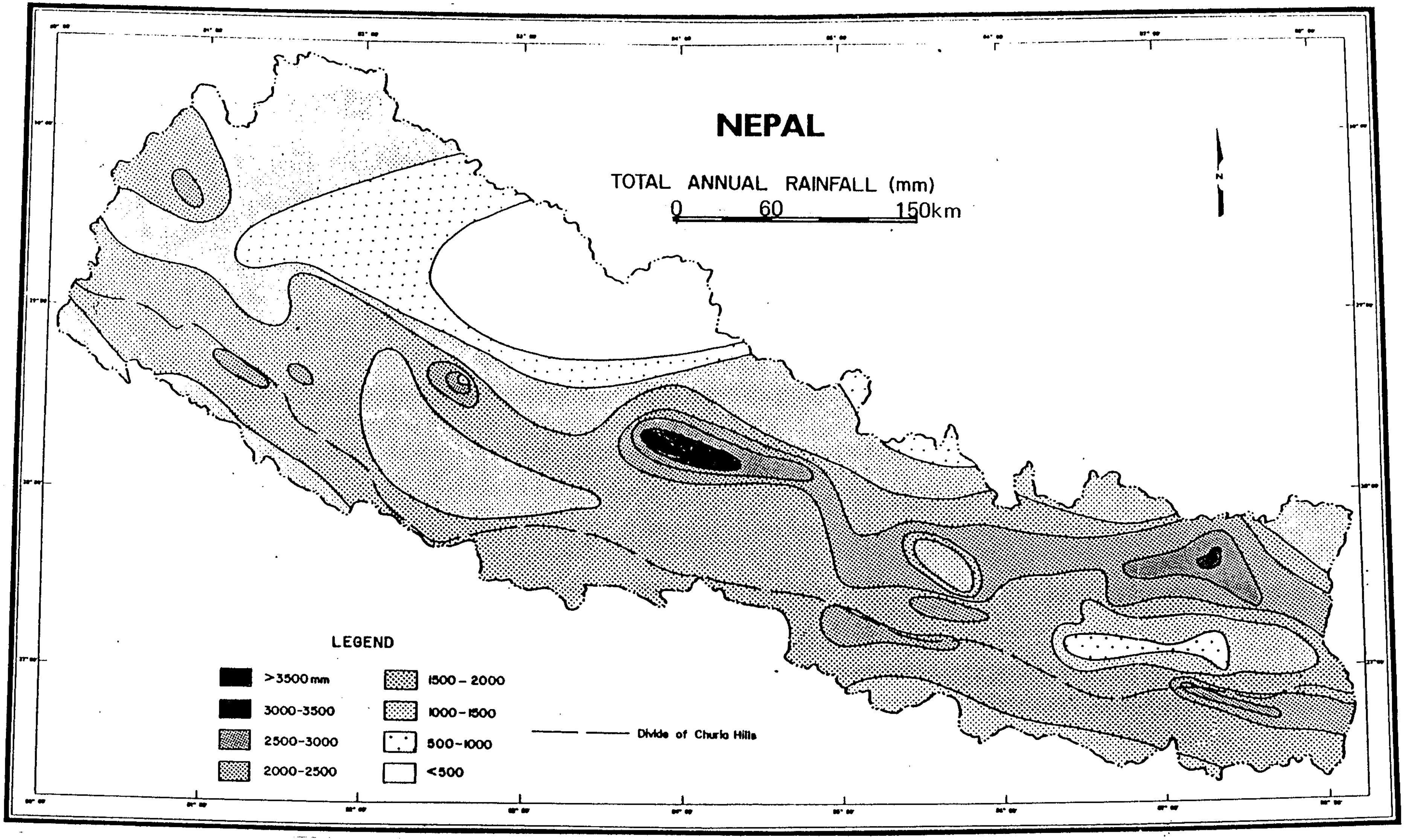


Figure 11 Mean annual precipitation (mm) in Nepal (after Chyurlia, 1984).

elevation on the basis of temperature lapse rates. The snowline was found at 2400 m in January, rising to about 5200 m in July.

Whereas it is quite feasible to monitor the areal extent of snow cover, it is much more difficult to estimate the actual amounts of snow falling on such areas. In the absence of snowpack surveys, Chyurlia (1984) combined information on seasonal variations in snow cover extent with mean monthly snowfall for a (very) limited number of high-elevation precipitation stations in Nepal, arriving at an annual total of ca. 1200 mm. Naturally, this must be considered as a crude first (under)-estimate.

The Geological Survey of India has conducted mass balance and other glaciological studies for several glaciers in the headwater area of the Ganges as well as in Sikkim since the mid-seventies.

As part of this research program a network of snow poles and ablation stakes was installed in the Beas basin in upper Himachal Pradesh (Anonymous, 1981a).

Bagchi (1982) used Landsat imagery to study the extent of snowcover in the same catchment. Given the difficulties associated with reliable estimation of snowfall at remote locations, the approach followed by Bagchi (1982) of linking fluctuations in snowcover extent with streamflow rates, seems to be more promising. After all, one of the principal applications of snow surveys is to determine the possible contribution of snowmelt to streamflow (cf. Section III.3).

Although annual precipitation totals are strongly dominated by the summer monsoon throughout the area (Figure 8), a certain amount of precipitation also falls in other months. In the western parts of the basin, 10-30% falls during the winter months and another 10-20% during the spring (or pre-monsoon) period.

Negligible quantities are recorded in October and November, the post-monsoon period (Dhar et al. 1987). Since most of the winter precipitation is generated by disturbances originating in the Mediterranean region, it is to be expected that such contributions decrease towards the east (Table 1).

Winter precipitation, mostly falling as snow, is especially important in the elevated semi-arid regions north of the main range. For example, at Kyelang in the Himachal Pradesh (total precipitation 555 mm/yr), winter precipitation constitutes some 45% of the total (Domroes, 1979), whilst at Jomosom (West-Central Nepal, average annual precipitation 255 mm), it contributes about 23% of the total (Chyurlia, 1984).

II.3.2 **Rainfall extremes**

In characterizing a location climatically, it is important to not only examine average rainfall totals and seasonal distribution, but also the frequency of extreme events, be it flood-generating rainfall or extended dry spells.

Chyurlia (1984) analyzed the year-to-year variability of annual and monthly precipitation for those stations in Nepal that had at least ten years of data. The result for annual precipitation, shown in Figure 12, suggests greater variation in the western parts of the country. This is due to the fact that towards the west the contribution of winter precipitation, which is far more variable than that of the summer monsoon (Chyurlia, 1984), becomes more important (Table 1).

Mooley & Parthasarathy (1983) examined above- and below-average annual rainfall extremes between 1871 and 1980 for 306 rainfall stations all over India, except for the northern mountainous districts. However, the Gangetic plain, as well as sub-Himalayan Bengal and Assam, were included in their analysis. They were unable to detect any trends or oscillations that were statistically significant. They concluded, therefore, that during the period of observation, annual rainfall

TABLE 1. Precipitation totals at selected stations in the Himalaya between December and March (after Dhar et al., 1987)

Station	Elevation (m)	Precipitation (cm)
Dalhousie, Himachal Pradesh	1960	58
Mussoorie, Garhwal Himalaya	2040	27
Mukteshwar, Kumaon Himalaya	2310	19
Jumla*, West Nepal	2300	13
Jiri*, East Nepal	2000	9
Darjeeling, North Bengal	2130	10

*Chyurlia (1984)

totals were distributed randomly in time.

In four years (1877, 1899, 1918 and 1972), more than 40% of India suffered a severe drought (over 70% in 1899!). Similarly, excessive rainfall was widespread over India in 1878, 1892, 1938 and 1961, with about 40% of the country being affected during the extreme event of 1892 (Mooley & Parthasarathy, 1983).

Interestingly, the occurrence of excessive droughts or rains in the northern districts that the present report is concerned with, showed comparatively little overlap between districts. In other words, droughts or excessive rains in western Uttar Pradesh would often show up in eastern U.P. as well, but not necessarily in Sub-Himalayan Bengal, and generally not at all in Assam (Mooley & Parthasarathy, 1983).

This would suggest that widespread flooding in the region is influenced by the areal extent of extreme rainfall more than by any other factor (Raghavendra, 1982; cf. section III.5).

The occurrence of a (very) wet or dry year seems to be related to the degree to which depressions are able to penetrate towards the west. Wetter years showed a distinctly higher proportion of depressions moving west of 80° E.L. (Mooley & Parthasarathy, 1983).

In addition, the regional distribution of rainfall appears to be strongly related to the location of the "monsoon trough", a zone of relatively low pres sure, normally running between southern Bengal and northwestern Rajasthan. The trough may shift towards the foothills of the Himalaya, producing a marked decrease in rainfall over India to the south of the trough, but a distinct increase over the Himalaya (Dhar et al., 1982a).

The synoptics of this situation, which is commonly referred to as a "break" in the monsoon, have been described by Ramaswamy (1962). To illustrate the magnitude of the phenomenon: Dhar et al. (1982a) reported about twice as much rainfall over eastern Nepal and Sikkim during "break days".

As for the extreme years, Mooley & Parthasarathy (1983) showed that during droughts (over Peninsular India) the average frequency of a "break" situation was about three times that observed during "flood" years. Also, the average length of the longest spell of "break" was two to three times higher during dry years.

Extreme amounts of rain falling over periods of one to several days are obviously of great practical significance because of their role in

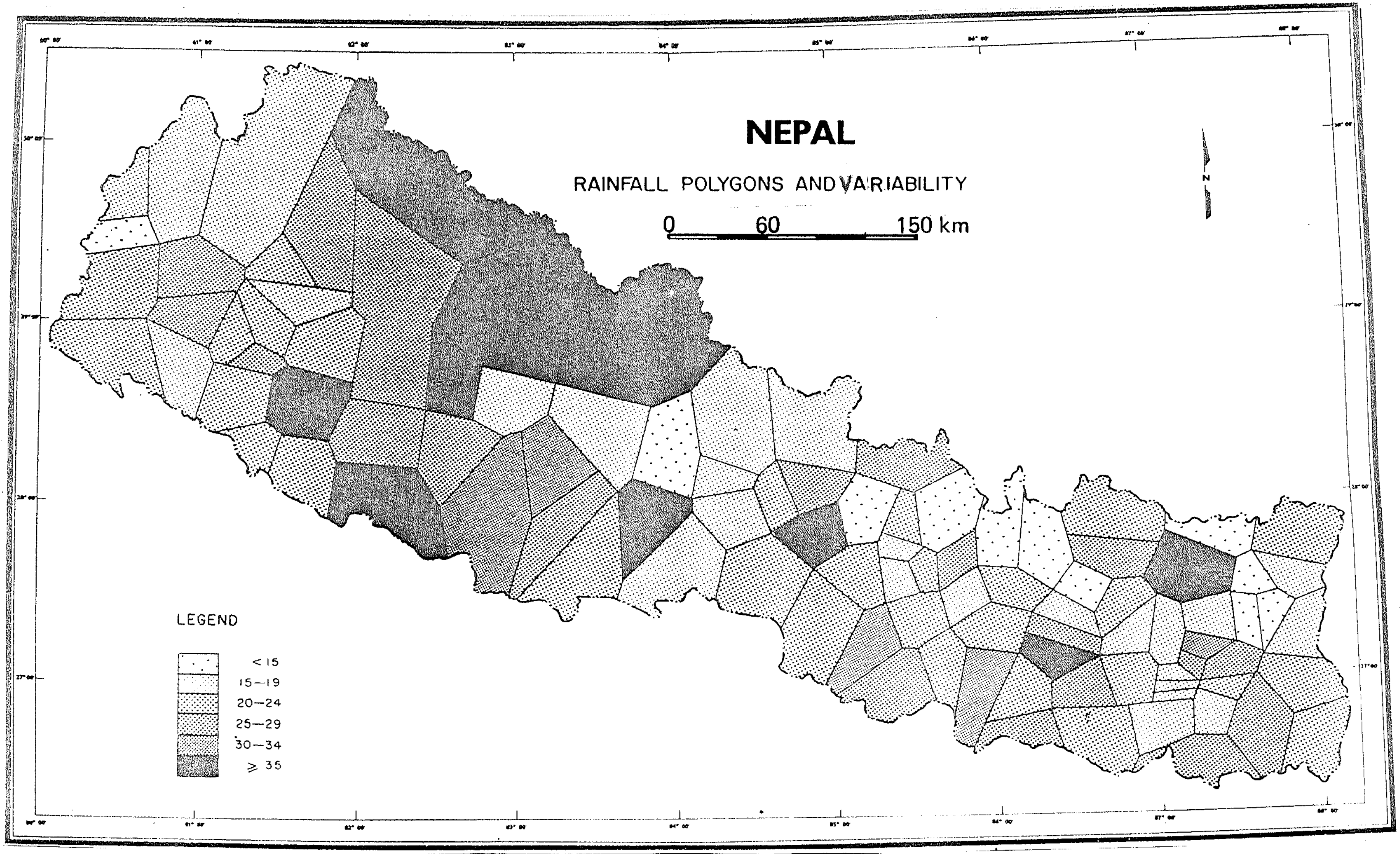


Figure 12 Variability index of rainfall polygons in Nepal (after Chyurlia, 1984).

generating major floods. This is especially so during the height of the summer monsoon, when soils all over the river basin are quite wet and have relatively little opportunity to accommodate such extreme additions of rain.

Although the issue of flooding will be worked out more fully in Sections III.4 and 5, the regional variation in extreme amounts of daily rainfall will be discussed briefly in the following.

The highest amounts of rainfall recorded on a single day at a number of selected stations throughout the basin are presented in Figure 13. Since the information was compiled from widely different sources, the periods of record differ between (groups of) stations, ranging from 30 to 60 years. Although this will doubtlessly introduce some bias, the regional trends should be reasonably representative.

By and large, maximum observed 24-hour rainfall figures exhibit a spatial pattern mirroring that of the annual or monsoonal rainfall (Figure 8). In other words, there is a trend of increasing values towards the east, and a decrease as one goes from the plains to the mountains (Figure 13).

As such, there again seems to be a natural tendency towards the greatest flooding potential in the eastern and lower parts of the basin.

The highest daily rainfall ever observed in the Ganges basin amounted to 823 mm, recorded at Nagina (Uttar Pradesh) in September 1880. The associated two-day total exceeded one metre of water, viz. 1042 mm (Sharma & Mathur, 1982).

The corresponding maximum daily rainfall in the Brahmaputra basin amounted to 1036 mm at Cherrapunji, (August, 1841; Holeman, 1968). It should be realized, however, that such extreme values often represent the core of a much larger "field" of rain with (much) lower amounts falling as one moves away from the centre (Figure 32).

Within the delta and adjoining coastal areas to the east, extreme rainfall may also be associated with typhoons. These occur, on average, about six times a year, arriving either in early summer (April, May), or in September-October, during which time much of Bangladesh is already inundated anyway.

Cyclones often generate waves that may be three to eight metres high, causing enormous damage, especially when they coincide with high tide (Haroun er Rashid, 1977).

III.3.3 **Evaporation**

In addition to rainfall, several other climatic parameters, such as temperature and humidity, solar radiation and windspeed, will affect the hydrological behaviour of an area of given geology, etc. The four parameters can be combined in a single variable, evaporation.

Eventually it is the balance between precipitation inputs and evaporation outputs, which determines the total amount of streamflow leaving an area. It is also a strong determinant of soil moisture status and, therefore, of a catchment's response to rainfall (Section III.4). As such it is important to investigate how the evaporative characteristics of a climate vary over a drainage basin, especially so in mountainous conditions with strong relief.

Arguably, of the many different techniques that are available for the estimation of reference evaporation rates, the one proposed by Penman (1956) is probably the most widely used among hydrologists. Although relatively data demanding, it is physically based and universally applicable (in contrast to the "rational" formula of Thornthwaite (1948), which is widely used in India, but strictly speaking not applicable outside northeastern USA).

Figure 14 presents the seasonal variation in monthly Penman reference evaporation over the Indian sub-continent. By far the highest values are recorded for the hot and dry pre-monsoon period (April, May). Evaporation during the summer monsoon, when the sun is at its highest, is

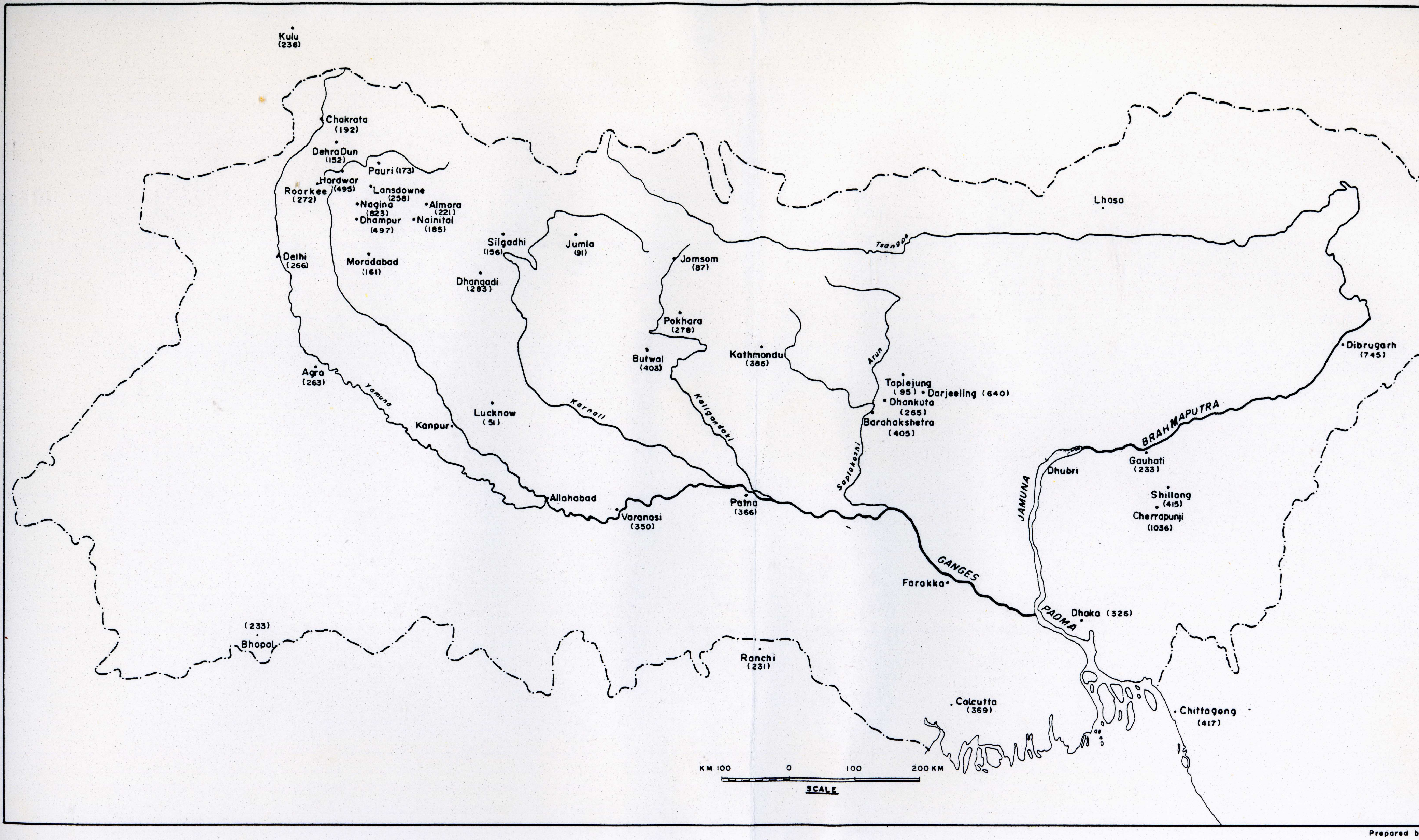
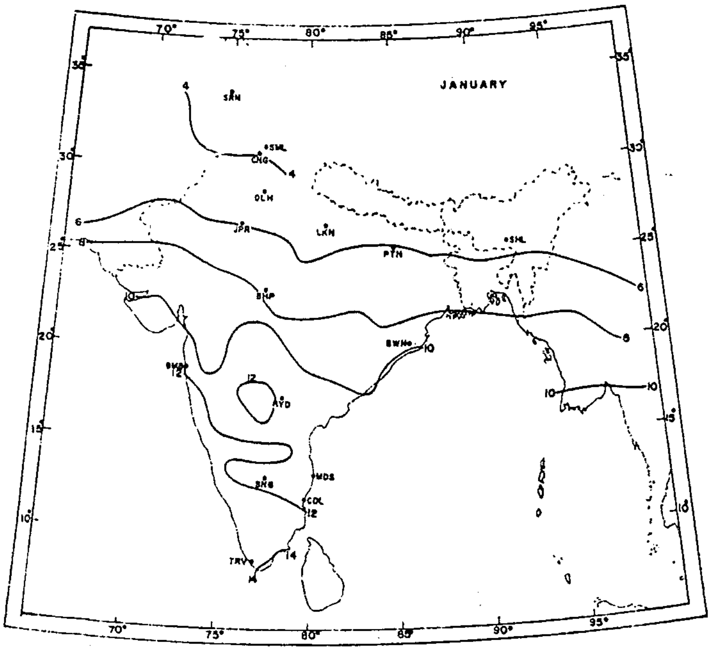
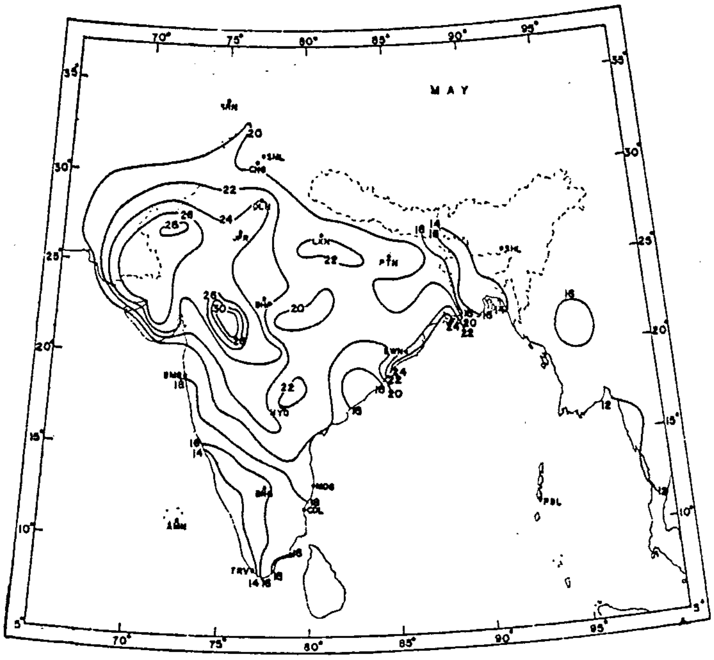
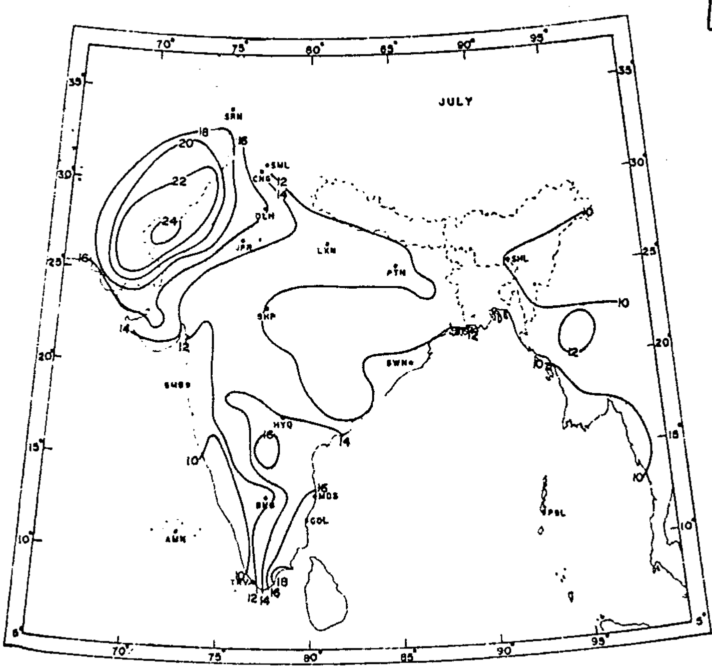


Figure 13. Approximate maximum observed 24-hr rainfall totals at selected stations in the Ganges-Brahmaputra River Basin (compiled from: Climatological Records of Nepal 1971-1986; Nayava, 1974; Raghavendra, 1982; Rao, 1981; Sharma & Mathur, 1982; Sharma et al., 1982; Starkel, 1972).







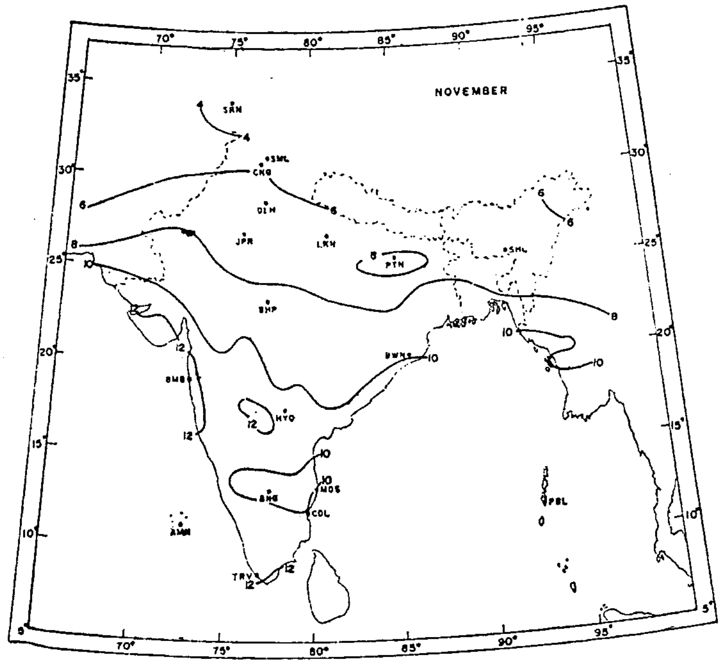


Figure 14.
Mean annual potential evapotranspiration according to Penman (cm) over the Indian subcontinent (after Rao, 1981).

moderated by increased cloudiness.

No such map has been published for Nepal, although Gurung & Lambert (1977) and Sapkota (1984) have reported values for individual stations in Nepal. Rather, the use of evaporation pans seems to be preferred in Nepal. Many of the figures of Penman evaporation given by Sapkota (1984) seem unrealistically high, and have therefore been omitted from the present report.

Sapkota obtained better results with an alternative model developed by Morton (1983), on the basis of which Figure 15 has been constructed. There is a reduction in annual evaporation as one goes north, that is, as one reaches higher elevations. Although this result was to be expected, it should be noted that Sapkota's estimates did not exhibit any clearcut relationship between annual evaporation totals and elevation.

Clearly, a characterization of spatial variations in evaporation rates in the Himalaya is still in its infancy.

II.4 VEGETATION AND LAND-USE

Vegetation is an overall expression of various environmental factors. Often, areas which are climatically similar will be characterized by similar forms of vegetation. As such, observations of vegetation distribution in the region could provide useful clues to identify environmentally homogeneous areas.

Given the variety in meso-climates and rock types found in the river basins, one expects an equally large range of vegetation types, some of which may exhibit strongly contrasting hydrological behaviour. In addition, heavy pressure on forests has resulted in the disappearance or degradation of the natural vegetation over time in many places (Plate 4), which will also have had an impact on existing hydrological patterns. Therefore, not only the spatial variations in vegetation and land-use types, but also the historical perspectives of "deforestation" need to be addressed.

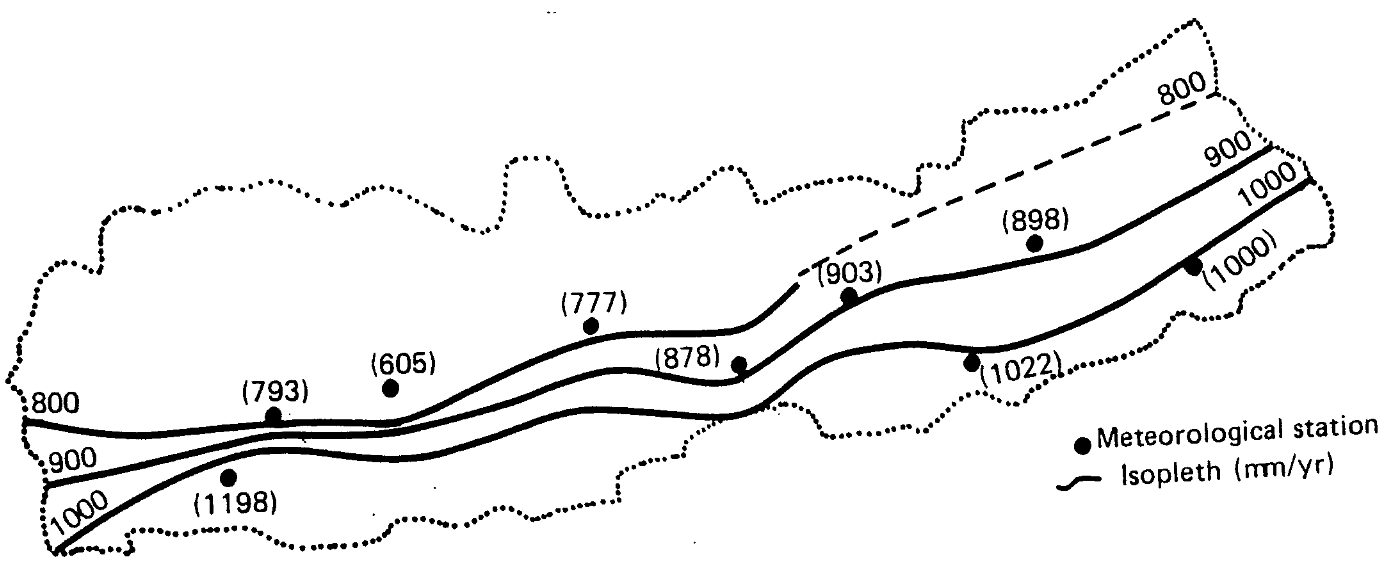


Figure 15 **Mean reference evapotranspiration (Morton, 1983) over Nepal (after Sapkota, 1984).**

II.4.1 Natural vegetation

With respect to the altitudinal zonation of vegetation in the Himalaya, Shrestha (1988) discussed several classification schemes proposed for the Nepalese Himalaya, and opted for the following simplified general sub-division:

Below 1000 m: Tropical zone
1000-2000 m:Sub-tropical zone
2000-3000 m:Temperate zone
3000-4000 m:Sub-alpine zone
Above 4000 m:Alpine zone

By and large, this sub-division is also valid for the Eastern and Kumaon Himalayas. However, as pointed out by Shrestha (1988), vegetation distribution is not only influenced by elevation (i.e. temperature) and slope aspect, but also by soil type and especially rainfall regime.

Whereas the Tropical zone of the Terai and foothills and the (Sub-)alpine zones of the High Himalaya show a more or less uniform character all over Nepal and further west, there are marked contrasts in forest type within the intermediate zones as one moves from east to west (Figure 16). Such contrasts assume extra importance, as it is within the Sub-tropical and Temperate zones that pressures on the forests are heaviest (Dobremez, 1976; Gupta, 1983; Mahat et al., 1986a, b). A north-south cross section is given in Figure 17.

Although the Southern Plateau and the Gangetic plain constitute the natural domain of various deciduous **tropical forest** types (containing *Shorea*, *Tectona*, *Dalbergia*, etc.), three millennia of human occupation have left very little of these forests, especially in the plains (Singh & Verma, 1987; Singh & Singh, 1987).

In the Bhabar zone (Figure 5), considerable tracks of deciduous tropical forest, generally dominated by *Shorea robusta*, can still be found (Plate 9), although the forest is being cleared at an extremely rapid rate in the Terai proper (Section II.4.2).

In general it can be stated that in the Terai (and probably in the entire region), wherever soil fertility and moisture status permit, there will be agricultural cropping, part of which is irrigated.

Although *Shorea* is the dominant tree species all over the Tropical zone, it is joined by a few co-dominants, which vary with changes in environmental conditions (Dobremez, 1976). The important thing in the present context is a tendency towards more open forest towards the west. Also, that part of the Siwalik hills bordering the Bhabar is extremely dry and vegetation response is that of a more open canopy and reduced stature. Smaller trees are also found in the *Shorea* forests of the Mahabharat near the transition to the Sub-tropical zone.

In the rainier eastern parts of the Ganges-Brahmaputra basin, the lowland forest initially still consists of *Shorea*, becoming (semi-)evergreen in the easternmost part of Assam. The whole of Assam valley contained about 13% of this type of forest in the mid-sixties (Das et al., 1987). The *Shorea* forests have been reported to have disappeared almost completely in the foothills of Bhutan (Sargent et al., 1985).

As for the **Sub-tropical zone** (1000-2000 m), an evergreen broad-leaved forest association, dominated by *Schima wallichii* and various chestnuts (*Castanopsis*) is found throughout the area between Assam in the east and the Kali Gandaki in West Nepal. As was the case in the Tropical zone, these forests become more species rich towards the east (Dobremez, 1976).

On drier southern slopes in Central Nepal and further east, the *Schima-Castanopsis* forest becomes more open and mixed with chir pines (*Pinus roxburghii*). Further west, chir assumes complete dominance (Figure 16).

Because the pine forests are generally intensively grazed and experience regular fires, they are believed by some to be a fire climax.

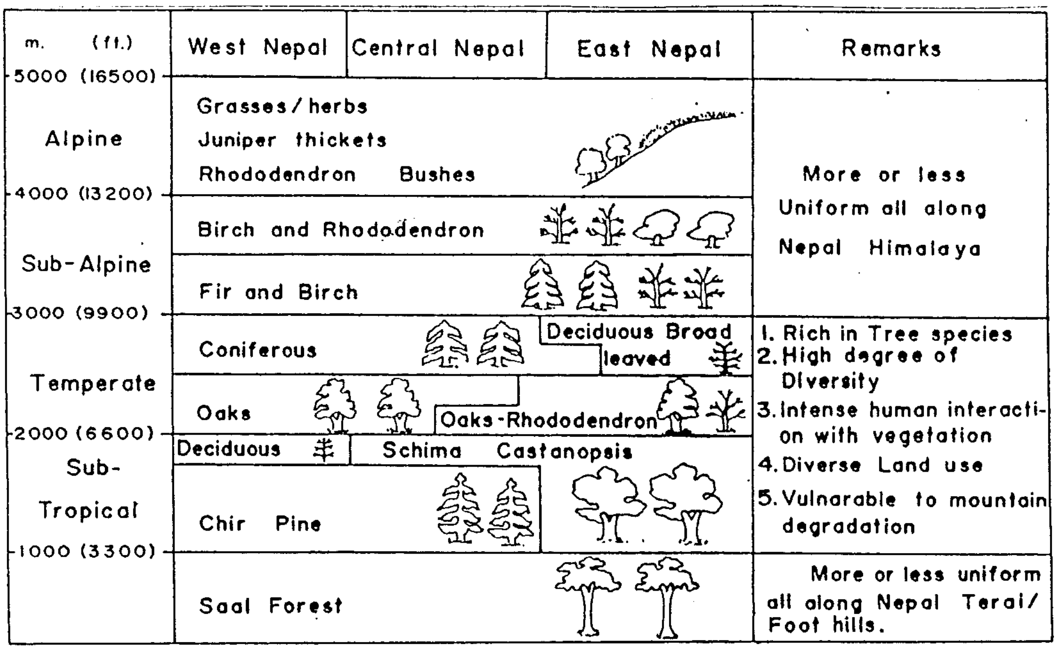


Figure 16 Vegetation zonation in the Nepalese Himalaya (after Shrestha, 1988).

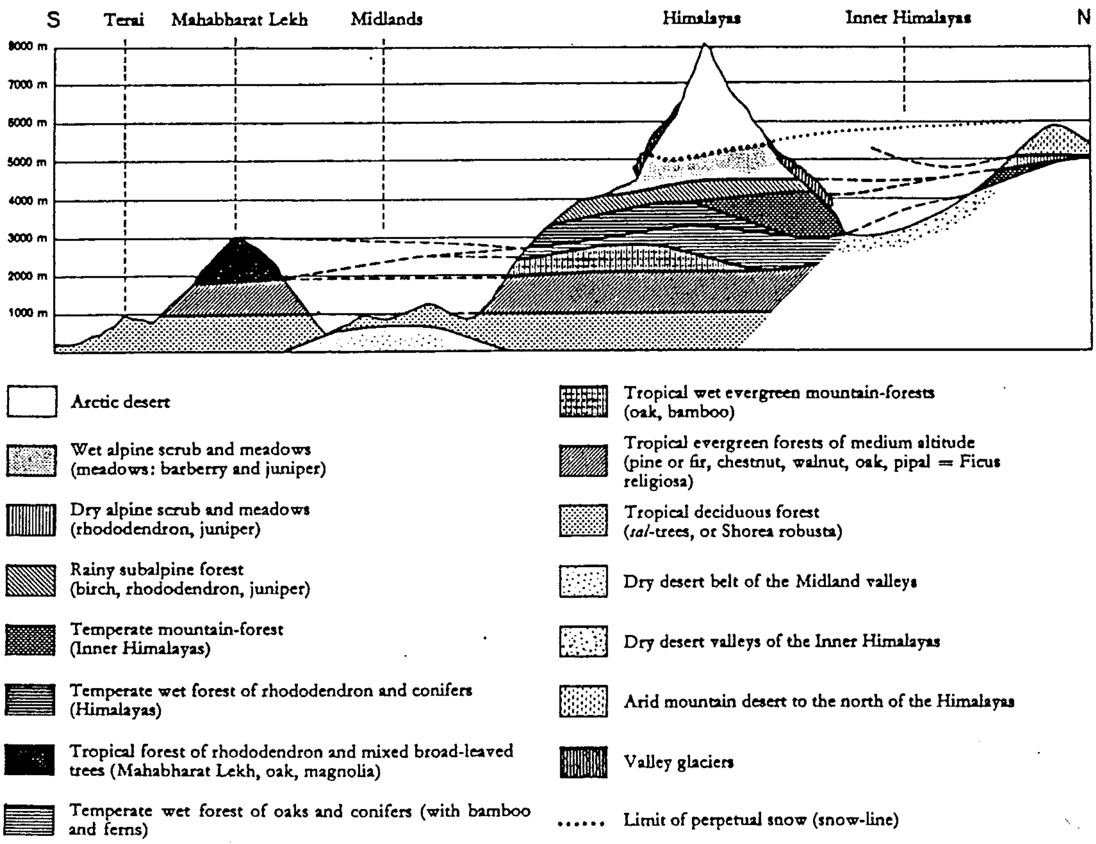


Figure 17 Cross-sectional representation of vegetation zones in the Himalaya (after Hagen, 1969).

Although this is true in Central and East Nepal (Shrestha, 1988), Dobremez (1976) reported the occurrence of pure stands of chir pine in West Nepal, which had not been subjected to fire or grazing as long as anyone could remember.

The sub-tropical broad-leaved forests have suffered heavy losses over the years along the entire area both quantitatively and qualitatively (see the next section).

Both in the Nepalese and Indian Himalayas, one may find stands with numerous ***Rhododendron arboreum*** trees, often in association with evergreen oaks, in the upper reaches of the Sub-tropical zone. Shrestha (1988) remarks that water courses are quite scarce in this forest type, which seems to prefer south-facing slopes.

In the Garhwal Himalaya, ***R. arboreum*** is mentioned as a typical associate of chir by Gupta (1983), whilst a similar situation has been described for Bhutan by Sargent et al. (1985).

The **Temperate zone** (2000-3000 m) is situated above the upper limit of widespread agriculture, the major limiting factors for cropping being high cloud incidence during the summer monsoon and low temperatures in winter (Shrestha, 1988).

The lower half of this zone is dominated by various evergreen species of oak, which constitute the main canopy, often attaining heights of 35-40 m. The lower strata are generally composed of laurels and magnolias. Especially in East Nepal and south of Annapurna Himal, the trees are heavily loaded with mosses (Plate 11).

Interestingly, the high humidity levels in these forests seem to reflect a large number of rainfall events rather than high rainfall totals in the case of East Nepal, and vice versa in the Annapurna region (Dobremez, 1976).

According to the same author, such differences in rainfall intensities also result in different mass movement hazards after clearing. As such, forest removal is supposedly less hazardous and therefore much more widely practised in East Nepal as compared to the mossy forest area in Central Nepal.

Since mossy, or "cloud forests" often receive additional inputs of moisture through the process of "cloud stripping" (Zadroga, 1981), their removal may have hydrological implications that differ from those associated with the clearing of "normal" forests (Chapter IV.1).

Lopping for fodder and grazing is a common practice in oak forests throughout Nepal and North India. In the East (Arunachal Pradesh, Assam), shifting cultivation (locally called ***Jhum***) is widespread (Prasad, 1987).

In East Nepal, this practice is relatively common between 1500 and 2300 m, with a cycle of only four to ten years. Bamboo is commonly found upon disturbance, as is a pioneer species called ***Eupatorium adenophorum***. Both species tend to hamper the regeneration of the oaks (Shrestha, 1988).

Towards the west, where drier conditions prevail, the protective undergrowth has often disappeared, largely as a result of intensive harvesting for fodder and fuelwood and regular burning to stimulate the fresh growth of grasses (Dobremez, 1976).

Here, the oaks are more drought resistant than their counterparts in the east. They represent the temperate equivalent to the sub-tropical chir belt below. From Nepal's Far West onwards, fairly open stands of deodar (***Cedrus deodara***) appear, becoming more common in the Kumaon and Garhwal Himalayas.

Above 2400-2700 m, where a snow cover is present for at least two months (Dobremez, 1976), the character of the oak forests changes. Whereas the main canopy is still dominated by oaks (mainly ***Quercus semecarpifolia***), several deciduous species (especially maple) come to the fore as well. Again, both the number of species and canopy density increases towards the east.

On the wettest sites (East Nepal, Sikkim), almost pure stands of rhododendrons replace the oaks. In the western parts of the upper Temperate

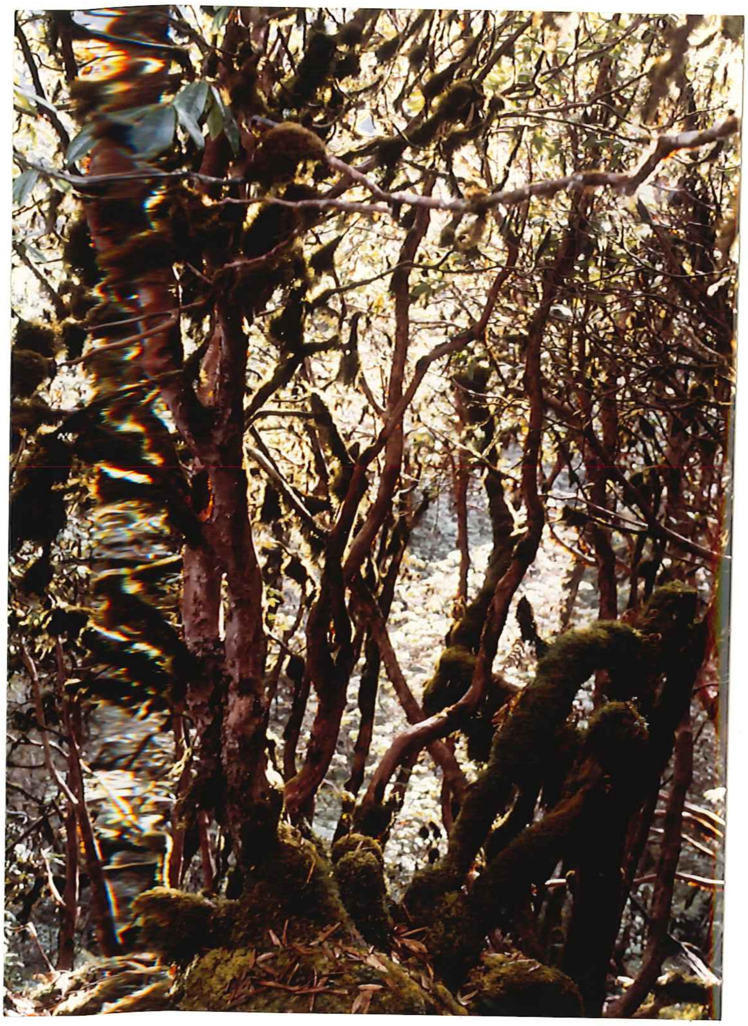


Plate 11

Mossy forest at 2600 m a.s.l. in Helambu, Central Nepal (photograph by R. Gerritsen).

zone, stands become more open and conifers increasingly important. Among the most common conifers are various pines, with junipers on the drier and fir on the colder sites. According to Dobremez (1976), abandoned fields in West Nepal at this elevation are often quickly invaded by *Pinus excelsa*.

Grazing pressure is high throughout the upper Temperate zone as high-altitude livestock moves down in search of warmth and low-altitude cattle migrate upwards in search of more fodder (Shrestha, 1988).

In the **Sub-alpine zone** (3000-4000 m) winters become much more severe and rooting opportunities are often restricted. This zone is dominated by relatively open stands of silver fir (*Abies spectabilis* and *Abies pindrow*), with oak and especially junipers on drier (western) and rhododendrons on wetter (eastern) sites.

At higher elevations, birch (*Betula utilis*) is quite common as well. One may also encounter groups of *Larix* (inner Himalayan valleys) or poplars and willows (valley bottoms) in this zone. The upper reaches of the Sub-alpine forests may be subject to disturbance by herders, who seek to enlarge the area for yak grazing (Dobremez, 1976; Sargent et al., 1985).

The vegetation becomes increasingly stunted with elevation, and above 4000 m (roughly coinciding with the timber line), in the **Alpine zone**, it is reduced to shrub size (junipers, rhododendrons), and finally consists

Plate 12 In the dry valleys north of the main range, the drier southerly slopes bear a steppe-like scrub vegetation, whilst the moister northern slopes are covered with a mixture of conifers (Manang district, Nepal).

of herbs and grasses only.

These grasslands are subject to grazing and collection of herbs for medicinal purposes (Shrestha, 1988). The shrubs may act as snowtraps. Diurnal variations in temperature and relative humidity are very large in the Alpine zone. Permanent snow is often found above about 5000 m (Dobremez, 1976).

Finally, a **Steppe zone** can be distinguished, characterized by open low vegetation dominated by *Caragana* sp. (Plate 12). It coincides with the arid Trans-Himalayan zone and its flora is strongly related to that of Tibet as expected (Dobremez, 1976). Interestingly, the permanent snow line is found at a higher elevation here than south of the Great Himalaya (Figure 17).

Diurnal fluctuations in temperature and especially relative humidity in the Steppe zone are considerably larger than for stations at similar elevations in the monsoonal part of the mountains (Dobremez, 1976).

II.4.2 **Agriculture**

The agricultural resource base in the Himalaya is severely limited by the steepness of the terrain, virtually all level land being restricted to alluvial landforms (Section II.2). As such, terracing of slopes is a dominant feature all over the region and has permitted farmers to grow crops on slopes that would have long since been washed away without such measures (Plates 5 and 19).

Cropping is done within the context of a mixed farming system, of which cattle and forests are also an integral part. Estimates of the number of hectares of forest land needed to support one hectare of cropland vary between three and six (Shrestha, 1988).

The link between the two forms of land use is constituted by livestock, which produces manure, draughting power, milk, etc. The animals are either allowed to graze in the forest (or its remnants), or are stallfed with fodder obtained from the forest. In addition, leaf litter is often collected for composting.

In general, there are two basic cropping systems in the mountains: one based on rice production on irrigated land, and another (more widespread) based on growing maize and millets on non-irrigated land. Potatoes often constitute the main winter crop.

There is an important difference in the types of terraces associated with the two systems: the ones used for irrigated agriculture must be flat (Plate 19), whereas the ones for rainfed cropping are often laterally or forward sloping (Plate 5).

As such considerable differences in surface erosion rates between the two types are to be expected (Section IV.2).

According to Shrestha (1988), yields per hectare of paddy, maize and millet have decreased by about 25% between 1970 and 1985 in the Arun area, a major river basin in East Nepal. This suggests that the present agricultural system is not fully sustainable (anymore). It also suggests that environmental degradation takes place both on the croplands and on the forest lands supporting them.

Coupled with an increase in population of about 1%/yr in the Arun area for the period 1971-1981 (Dunsmore, 1988), a rather grim picture emerges (Hrabovsky & Miyan, 1987).

An even less sustainable agricultural system (shifting cultivation) is practiced in the (sub) tropical zones of the eastern half of the mountains. Locally known as *Khorea* in East Nepal, as *Tsheri* in Bhutan, and as *Jhum* in North-east India, this type of farming is often found at (very) steep slopes, whereas the cycle of rotation has decreased to less than five years in many occasions (Shrestha, 1988; Toky & Ramakrishnan, 1981). Such short cycles have been shown to be not only economically non-viable, but also highly detrimental to the environment in the case of North-east India (Mishra & Ramakrishnan, 1983a,b).

In Nepal, abandoned fields are often quickly colonized by ***Eupatorium adenophorum***, initiating the natural succession to the original forest. However, due to the increasingly shorter cycles, a mixture of young secondary forest types occur, which do not get the chance to mature. ***Eupatorium*** also invades marginal grazing land (Shrestha, 1988). In the Sub-alpine zone sheep, goat and yak graze the pasture lands, but herds are maintained only in the vicinity of forest (Shrestha, 1988). Grazing has always been important above 3000 m and the high pressure on the oak forests of the Temperate zone, both from above and from below, has already been indicated. Table 2 illustrates this rather well for Nepal.

By contrast, in the plains, the bulk of the land is devoted to agricultural cropping, much of which is irrigated. Whereas wheat is the main crop in the upper parts of the Gangetic plain, rice is becoming more important as one goes east (Singh & Singh, 1987; Singh & Verma, 1987).

An interesting observation on cropping patterns in the lowlands of Bangladesh has been reported by Currey (1984). The traditional system (***aus*** crop) was adapted to the annual inundation in that the crops were harvested before mid July, just before flooding would normally set in. Modern agriculture, on the other hand, involves year-round cropping and is consequently much more vulnerable to flooding (Figure 18; see also Paul, 1984).

II.4.3 Forest conversion and degradation: a historical perspective

Much has been made of accelerated deforestation in the Himalayas since the last three decades or so, and of the ensuing environmental consequences, both locally and downstream (Eckholm, 1975; Bowonder, 1982; Myers, 1986).

This rapid rate of deforestation

Table 2. Relative proportions of land-use types in the physiographic regions of Nepal (%) (after Carson et al., 1986)

	Irrigated Rice	Rainfed Cultivated	Grazing and Scrub	Degraded Forest	Closed Forest	Other*	Total Percent	Total 000ha
Terai	50	6	4	2	26	12	100	2110
Siwaliks	7	5	7	9	68	4	100	1886
Middle Mountains	7	15	36	18	22	2	100	4443
High Mountains	1	5	30	15	40	8	100	2959
High Himal	0	0.0**	29	2	2	67	100	3349
TOTAL								14748

*boulder, rock, ice etc.
**0.0 less than .5%

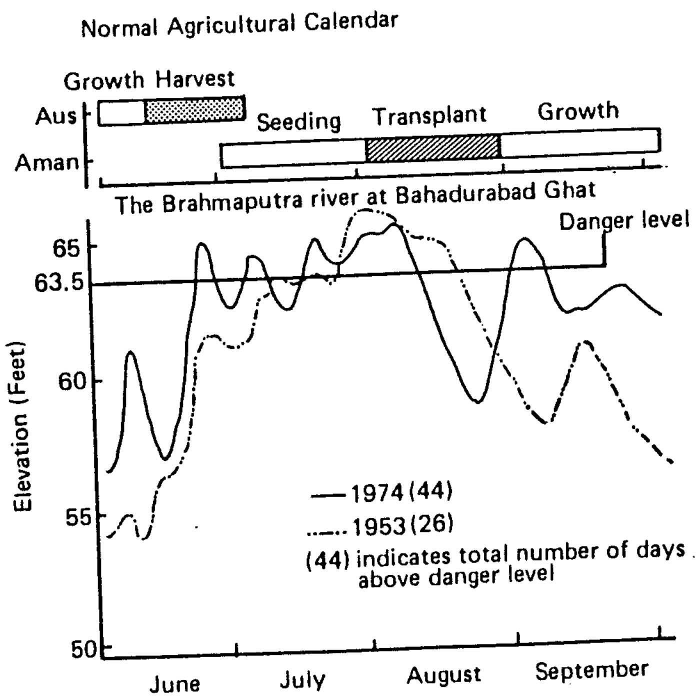


Figure 18 Agricultural calendar in relation to timing of flooding in Bangladesh (after Currey, 1984).

is often seen as a recent, yet terminal phenomenon, unless truly drastic measures are taken. It has been suggested, for example, that there would be no more accessible forest left anymore in most of Nepal around the year 2000 if the presently assumed rates were to continue (World Bank (1980) in Gilmour, 1988).

However, there is a growing body of literature, mostly based on the interpretation of time sequences of aerial photographs or satellite imagery, which indicates that there has been very little change in the area occupied by forest and agricultural land in Nepal's Middle Hills since 1964 (Bajracharya, 1983; Strebal, 1985; Malla, 1985).

In addition, evidence based on oral history research in two hill districts in central Nepal suggested that the forest boundaries have remained more or less stable for at least a century (Mahat et al., 1987).

In other words, the bulk of deforestation in the hills must have taken place well before the present century (Mahat et al., 1986a, b).

In the Terai, on the other hand, rates of forest clearance for agricultural and other purposes have been extremely high over the past few decades, even though the area has known a long history of exploitation (Tucker, 1987). Malla (1985) reported that between 1954 and 1981 more than 60% of the tropical forest of the Nepalese Terai had disappeared. The rate of clearance between 1977 and 1981 was as high as 3%/yr.

The Land Resources Mapping Project estimated existing forest resources in the Middle and High Himalayas of Nepal as per 1984 at 42 and 35% respectively, and at 23% in the Terai (Malla, 1985).

Whereas the *area* occupied by "forest" land may not be subject to great changes, there is reason enough for grave concern. Although estimates of demand for fodder and fuelwood vary widely (Thompson & Warburton, 1985), there can be little doubt that in many areas demands exceed supplies (Kayasthra, 1988).

The decline in crop yields in East Nepal referred to in the last section may be seen as a sign on the wall in this respect (Shrestha, 1988). All over the hill region, there has been a marked reduction in forest density, especially on the edges bordering agricultural lands.

As such, deforestation is not so much a loss of forest area, but rather a loss of forest quality (Panday, 1986). Moench & Bandyopadhyay (1986) have described some of the mechanisms involved, showing how a loss of forest cover can occur even though overall biomass productivity far exceeds village demand (see also Gilmour, 1987).

Despite increasingly successful attempts at reforestation and a growing interest of farmers to plant trees on their own lands, the balance is still heavily on the negative side (Gilmour, 1988).

Nevertheless, there is a growing awareness of the situation at all levels and there is reason to believe that the predicted mega-crisis may well be far less terminal than thought by some (Gilmour, 1988; Panday, 1986).

As for the Indian Himalaya, estimates of the areal extent of forest land in the Hill Districts of Uttar Pradesh, sub-Himalayan West Bengal and Sikkim, suggest a situation similar to that of the Middle Himalaya of Nepal, with values between 30 and 40% of the total area (Tejwani, 1985). These figures, however, need to be looked at with a fair deal of caution. For example, Gupta (1983) showed that of the 3,253,000 hectares of "forest land" in the Hill Districts of Uttar Pradesh, 15% consisted of ice, rock and alpine grasslands, and another 27% of severely degraded scrub and wasteland not under the management of the State Forest Department.

In addition, there is no guarantee that the area actually managed by the Forest Department does consist of well-stocked forest. Tejwani (1985) estimated that about 60% of the land managed by the Department was considered "exploitable".

According to Tucker (1987), the first wave of massive deforestation in the Indian Himalaya occurred in the 1850's and 1860's in the wake of the establishment of British control in the upper Ganges plains and the associated building of a railway network. Completely unregulated exploitation took place, especially in the Tropical (***Shorea robusta***) and Sub-tropical (Deodar) zones (cf. Section II.4.1).

With the establishment of the Indian Forest Service soon after, the damage to ***Shorea*** forests (not deodar) was repaired to a fair degree and for more than a century a sustained yield production system was effected.

However, a considerable portion of the forests were placed under the control of the Revenue Department, which did not do much to preserve them. In addition, private contractors caring little for environmental values, were generally involved in harvesting the timber for distant markets until fairly recently. Meanwhile, access to these forests by the local population was limited (Tucker, 1987), which led to tensions long before the Chipko movement came into existence. Under pressure of the latter, the Indian Government has recently imposed a ban on all commercial forestry operations in the Himalaya, initially for fifteen years.

Degradation of the forest continues, nevertheless, as a result of ongoing exploitation by the local population (Moench & Bandyopadhyay, 1986).

As for the rate of "deforestation" in the Indian Himalayas, Tejwani (1985) suggested a similar trend as for the entire nation, which experienced an overall loss of forest over the period 1951 to 1980 that exceeded planting rate by some 16%. This may well be a conservative estimate in the light of the findings of Gilmour (1988) for the Nepalese Middle Hills.

In Bhutan, Arunachal Pradesh and Purvanchal, forest cover is much better than in the West (over 60%; Tejwani, 1985), although shifting cultivation is widespread. According to Sargent et al. (1985), Bhutan enjoyed a forest cover of about 55 % in 1978, 22 % of which was close-canopied and almost certainly primary. The destruction of the ***Shorea*** forests of Bhutan has already been indicated, but Sargent et al. (1985) also signalled heavy pressures in the Sub-tropical vegetation zone, mainly in the form of (commercial) logging and grazing. Shifting cultivation is especially a problem in the south-eastern parts of Bhutan (Upadhyay, 1987).

In Meghalaya and Nagaland, forests make up only 10 to 20% of the land area and the ***Jhum*** cycle has become critically short (Section II.4.2).

III. HYDROLOGY AND SEDIMENTATION

III.1 ANNUAL WATER YIELD

In analyzing highland-lowland interactions in the Ganges-Brahmaputra river basin, it is obviously of interest to examine what proportion of the flow carried annually by the two main rivers originates in the two major upland units, viz. the old plateaus of relatively low relief in the south and the much steeper and more elevated Himalaya in the north.

Rao (1975) presented annual streamflow totals for India's major rivers, suggesting that 30% of the flow arriving at the Farakka barrage in the Ganges is contributed by the southern tributaries, two-thirds of which is delivered by the Yamuna alone (Figure 1).

An inspection of streamflow records for the major northern tributaries flowing through Nepal (summarized by Alford, 1987) indicates that of the remaining 70% of the flow at Farakka, ca. 45% (i.e. 30% of the total) comes from Nepal.

Similarly, the Himalayan tributaries supply about 63% of the amount of water discharged annually by the Brahmaputra at Bahadurabad in Bangladesh (Figure 1). About half the Himalayan contribution is made up by the Brahmaputra (then called Dihang) itself at the point where it emerges from the mountain range (Sharma, 1985). It can be safely assumed, therefore, that about two-thirds of the total flow of the combined river system originates in the Himalayas and the adjacent plains.

The vast amounts of water that are discharged each year into the Bay of Bengal ultimately derive from parts of the river basin showing important climatic contrasts. As such, major spatial variations in streamflow amounts and timing are to be expected.

At the bottom end of the range one finds such rivers as the Chambal in the far western corner of the basin (140,000 km^2), and the upper Brahmaputra (Tsangpo, 153,200 km^2) in western Tibet, which show runoff totals of only 225 and 200 mm/yr respectively (Rao, 1975; Guan & Chen, 1981), reflecting the arid conditions prevailing over much of their drainage basins.

At the other end of the scale, there are rivers such as the Burhi-Dihing (6000 km^2) in Assam, or the Seti, Chepe and Balephi rivers (300-600 km^2) in the High Himalaya of Central Nepal, all of which have streamflow totals over 2500 mm/yr (Sarma, 1986; Alford, 1988a).

In contrast to the Burhi-Dihing, which is entirely rainfed and of relatively low elevation, these high-altitude Nepalese rivers receive yet unspecified contributions of snowmelt.

Rivers draining the hot southern plateau area, such as the Betwa, Ken and the Son, typically exhibit streamflow values of 350-450 mm/yr (Rao, 1975).

As already indicated, the Himalayan tributaries show a wide variation, reflecting an equally wide range of hydrometeorological conditions. For example, within the Sapt Kosi river basin in eastern Nepal, the discharges of the major tributaries range from 580 mm/yr for the Arun to 1920 mm/yr for the adjacent Tamur (Alford, 1988a).

This can be explained by the fact that the Tamur river originates on the south-facing slopes of the High Himalaya, which experience relatively high precipitation totals (Figure 11). In addition, the Tamur must receive important contributions in the form of meltwater from the glaciers of Mt. Kanchenjunga.

The Arun, on the other hand, has about 90% of its drainage area located north of the main range in the Trans-Himalaya, where both precipitation and snowmelt contributions are modest (Guan & Chen, 1981).

In general, the larger a watershed's proportion in the rainshadow zone north of the Great Himalaya, the lower will be the annual water yield

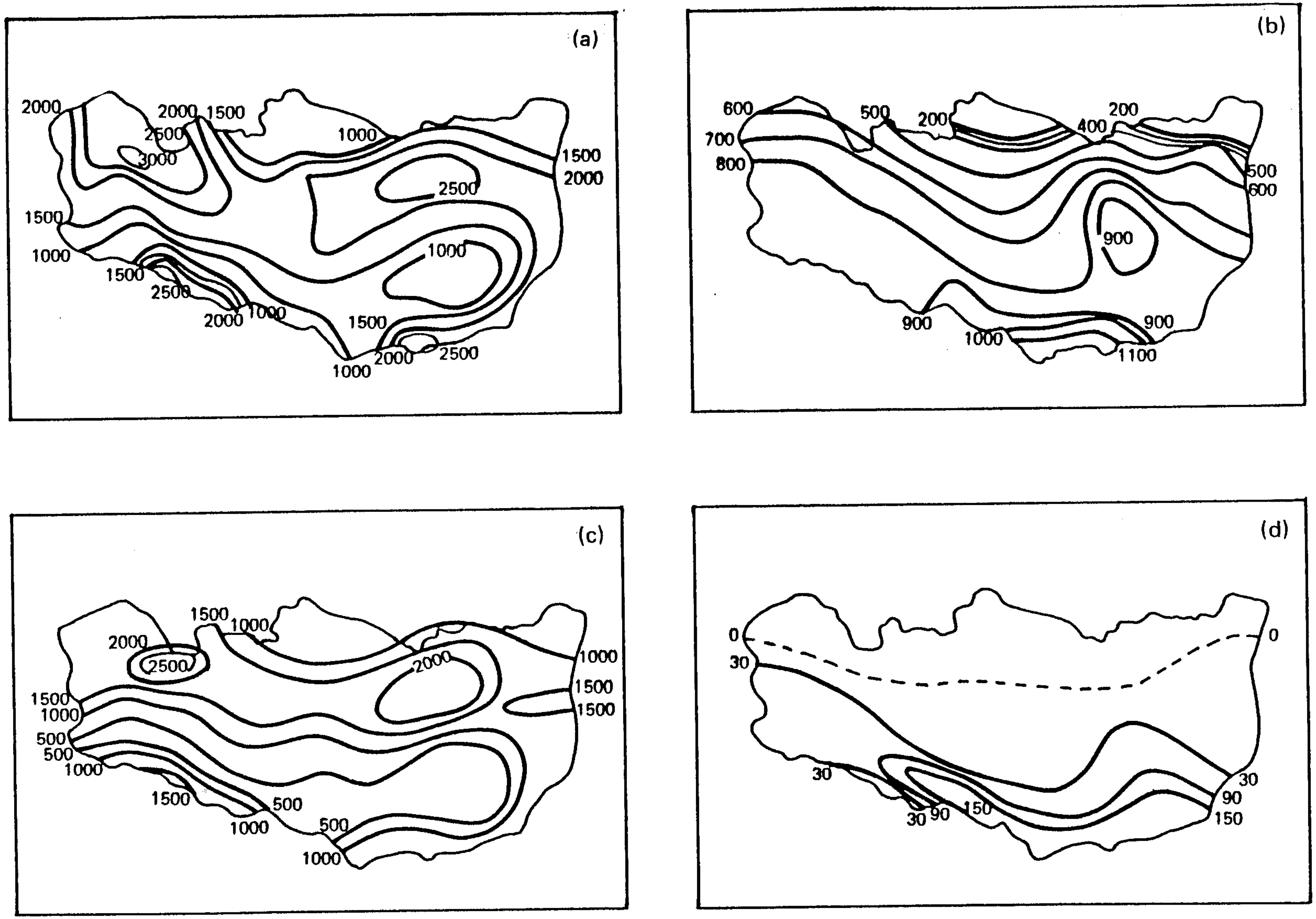


Figure 19. (a) Average precipitation, (b) actual evapotranspiration, (c) surplus, and (d) water deficiency (mm/yr) in the Nepalese part of the Sapt Kosi River Basin (after Subramanyam & Upadhyay, 1983).

(Alford, 1988a).

Another group of rivers originate in the Middle Himalaya. Found at intermediate elevations (mostly between 500 and 2500 m a.s.l.), these rivers experience a strong monsoonal influence and relatively high evaporative demands and no significant snowmelt. As such their streamflow totals may be expected to lie in between the above-mentioned extremes.

However, variations within this physiographic zone are especially large as a result of rain-shadow and orographic effects (Figure 11).

Geological factors, such as leaky river beds (e.g. in the Siwalik-Dun zone (Chyurlia, 1984) or in limestone areas (Rawat, 1985) may also play a role. Typical examples are the Bagmati and East Rapti rivers in Central Nepal (850 and 1640 mm/yr respectively), the Sarada and West Rapti in West Nepal (750 and 690 mm/yr respectively), and the Bemunda in Tehri Garhwal (1200 mm/yr) (Sharma, 1977; Puri et al., 1982; Chyurlia, 1984; Alford, 1988a).

Each of the larger Himalayan river systems represents a different combination of the three major environments, i.e. the dry Trans-Himalaya, the icy Great Himalaya and the monsoonal Middle Himalaya. As such, each of these large basins exhibits a unique, although to some extent predictable, streamflow total (Chyurlia, 1984).

An analysis of regional streamflow patterns in Nepal is underway and should become available in the near future (D. Judge, personal communication).

III.2 WATER BUDGETS

The spatial and temporal variability of water in any environment is determined by the relationships between inputs (generally as rain or snow), storages (in soils, geological deposits, lakes, glaciers or a seasonal snowpack), and outputs (streamflow, evaporation to the atmosphere).

Their interaction can be described conveniently with the help of the water budget equation, which is essentially a continuity equation (Ward, 1975):

$$P = Q + ET \pm \delta S \pm L \quad (1)$$

when P = precipitation
Q = streamflow
ET = evaporation
δS = changes in stored amounts of water/snow
and L = subterranean gains or losses,

with all elements expressed as mm/time unit (e.g. month, year).

Often this basic equation is simplified by computing the balance for an entire seasonal cycle ("water year"), which tends to reduce the change in the storage term to zero (Ward, 1975).

In addition, careful selection of river gauging sites with respect to geological conditions may minimize the sub-terranean term as well. For such a "watertight" drainage basin equation (1) simplifies to

$$P = Q + ET \quad (2)$$

with the components expressed as mm/yr.

It follows, at least in theory, that first approximations of streamflow totals can be made for ungauged rivers from an adequate knowledge of the spatial variations in precipitation inputs and evaporation outputs.

Such an approach meets with considerable difficulties, however, in the extremely rugged terrain of the Himalayas.

As already indicated in the sections on climate, there is a dearth of knowledge regarding the amounts of precipitation falling at higher elevations and about evaporation rates in general.

The general underestimation of precipitation inputs is illustrated rather well by the unrealistically small differences between annual precipitation and streamflow totals (i.e. ET) for several large Nepalese river basins as computed by Chyurlia

(1984).

According to these figures, the areally weighted average evaporation over the Sapt Kosi basin (57,200 km^2) would amount to only 105 mm/yr. Corresponding values for the Kali Gandaki (34,440 km^2) and Karnali (19,260 km^2) basins read 135 and 190 mm/yr respectively (Chyurlia, 1984).

More plausible, but essentially untested, are the results of the water balance computations by Subramanyam & Upadhyay (1983) and Alford (1987) for mountainous catchments in Nepal. The former applied the classical Thornthwaite & Mather (1957) approach to compute evaporation rates from temperature data and compared these with precipitation data to define zones of water surplus and deficiency in the Sapt Kosi basin (Figure 19).

Alford (1987) took the procedure one step further by manipulating the precipitation and evaporation relationships with altitude in such a way, that the computed streamflow outputs (cf. equation 2) for his study basin - the Seti Khola near Pokhara in West Nepal - matched the measured amounts of flow.

Naturally, the question as to what extent this apparent success in simulating streamflow totals, using a relatively simple procedure, is the result of a fortuitous combination of variables, remains unanswered. This would involve measurements of the various components of the water budget for carefully selected catchment areas in the respective physiographic zones.

One Himalayan catchment, for which such data have been published, is the 1754-ha Bemunda watershed, situated between 800 and 2200 m in the "Middle Hills" of Tehri Garhwal (Puri et al., 1982). Over the period June 1981 to May 1982, some 1210 mm of streamflow were recorded, against a rainfall total of 2235 mm. This would imply an evaporation figure of 1025 mm/yr (including δS) for this largely forested catchment.

Unfortunately, Puri et al. (1982) considered their streamflow figures underestimates, because of yet unknown amount of groundwater flow underneath their measuring flume. Clearly, the study of water budgets in the Himalayan context has only been touched upon until now and much more work is urgently needed if we are to increase our understanding of the hydrological behaviour of this environment (Kattelmann, 1987).

More information in this respect is available for the rivers draining the southern plateau area. Annual totals of precipitation and streamflow, as well as evaporation increase as one goes from west (Chambal river) to east (Son, Godavari).

Typical values for ET range from 530 mm/yr in the west (mean rainfall 755 mm/yr; Jha et al., 1988) to 890 mm/yr in the south-east (mean rainfall 1185 mm/yr; Biksham & Subramanian, 1988).

III.3 SEASONAL VARIATIONS IN STREAMFLOW

In a strongly seasonal climate like that of the Ganges-Brahmaputra river basin, monthly variations in streamflow assume great importance. Mean monthly flows for several basins of intermediate size, considered representative for the major physiographic zones within the Himalaya described earlier, are given in Figure 20.

Although all four basins exhibit maximum flows during the summer months, when both rainfall inputs and rates of snowmelt are at a maximum, there are marked differences with respect to the amplitudes in seasonal variations.

Absolute variations are small in the Tibetan example (Figure 20a), where streamflow is the result of a combination of groundwater outflow from alluvial deposits (50%) and modest contributions of meltwater (20%) and rainfall (30%) in the summer months (Guan & Chen, 1981).

The groundwater contribution typically decreases in spring, when neither melt nor rainfall replenishes the system (Figure 20a). Incidently, no such decline in flow occurs in larger Tibetan basins, reflecting the

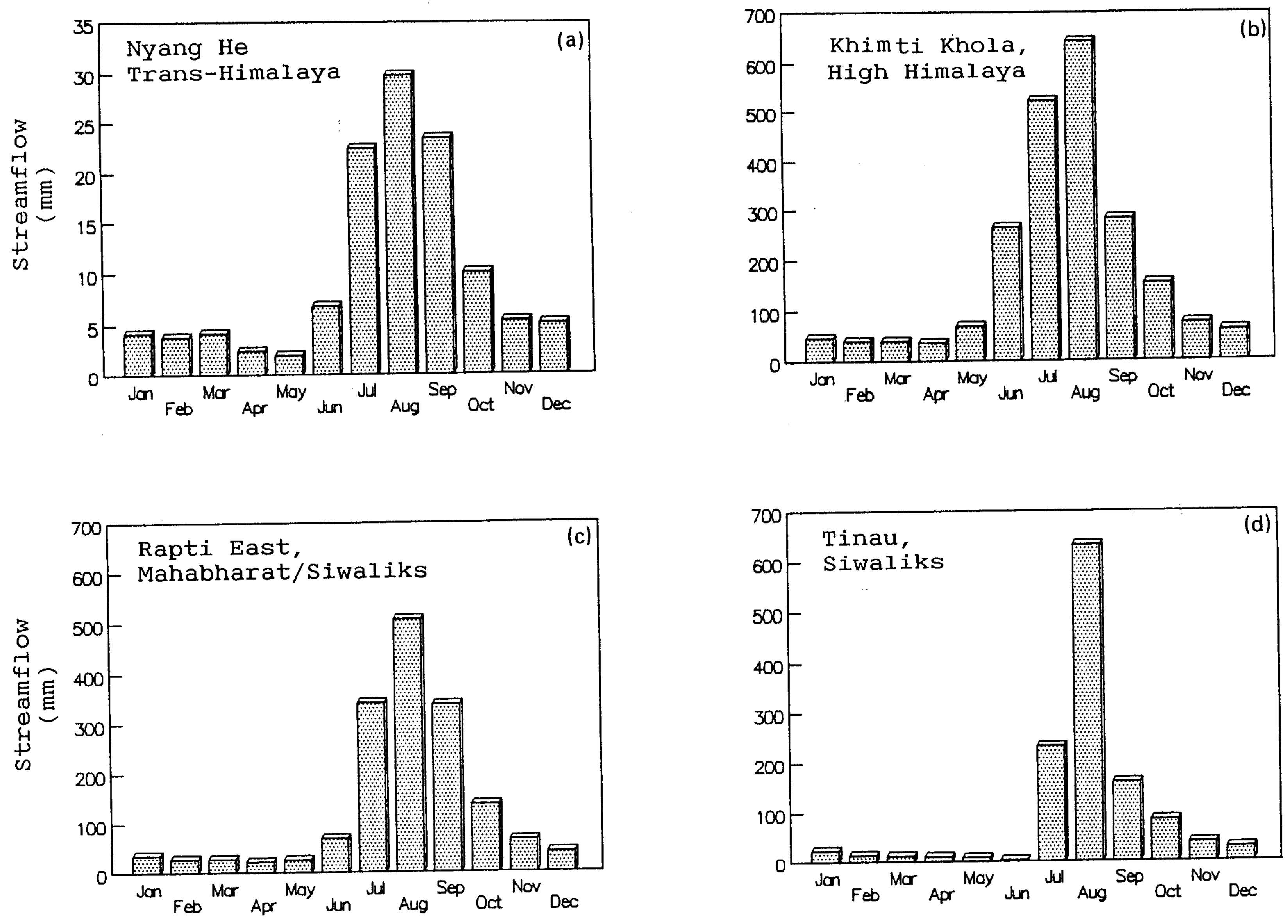


Figure 20 Monthly streamflow (mm) for selected Himalayan rivers (based on data presented by Guan & Chen, 1981; Surface Water Records of Nepal; Sharma, 1977).

much higher groundwater reserves in large river valleys (Guan & Chen, 1981).

Since the Nyang He watershed is about ten times as large as the other catchments in Figure 20, strictly speaking its flow regime cannot be compared directly. Presumably, springtime recession would almost certainly have been steeper and the ratio between maximum and minimum flows correspondingly larger.

The flow regime of the Khimti Khola, which rises not far from the source of the Nyang He on the south-facing slopes of the main range, seems much more extreme at first sight (Figure 20b).

This is mainly a matter of the amounts of water involved as the ratios between maximum and minimum flows are quite comparable for the two basins (18 vs. 16 respectively).

Indeed, the High Himalayan zone appears to be the richest area of water (Alford, 1987) as a result of a combination of factors. These include the full exposure to moist air of the summer monsoon, relatively low evaporation rates associated with high elevations (Subrahmanyam & Upadhyay, 1983), and sustained meltwater contributions in spring (Figure 20b). This moderating effect of snowmelt contributions on low flows in the pre-monsoon months is lacking in the case of streams originating in the Middle Hills and the Siwaliks (Figure 20c, d).

In addition, rainfall is more intense here as compared to the High Himalaya (Chyurlia, 1984). Therefore, the ratios between maximum (Q_{max}) and minimum (Q_{min}) monthly flows tend to increase substantially under these conditions, i.e. the streams become more "flashy". This is especially so for streams rising in the southern Mahabharat-Siwalik zone, where strong rainfall leads to high summer flows, whilst the highly permeable nature of the valley fills favours leakage into the river beds, thus further reducing baseflows in winter and spring (Figure 20d; Plate 7 & 16).

Ratios between Q_{max} and Q_{min} are also seen to increase as one goes from east to west within a physiographic zone, reflecting the corresponding trend in rainfall seasonality (Sharma, 1977).

A semi-quantitative interpretation of the flow regimes of several large Nepalese river basins with respect to the origin of the water was undertaken by Chyurlia (1984).

He distinguished four major flow components, viz. inflow of groundwater from valley deposits (baseflow Q_b), meltwater (effectively between February and August only, and peaking in June: Q_s), "interflow" (Q_i) and "direct flow" (Q_d) (Figure 21).

In this view, "interflow" is considered to represent the portion of the precipitation, that is temporarily held in storage in the soil mantle before being released to the streams. "Direct flow", on the other hand, enters the stream directly (presumably during rainstorms) without being stored.

Although Chyurlia (1984) himself considered the separation between Q_i and Q_d "a rather arbitrary affair in the absence of detailed basin studies", it would seem as though this contention applies to the rest of the separation lines in Figure 21 as well.

For example, a glance at Figure 22, depicting the seasonal course of meltwater discharge from a glaciated basin in the Khumbu area of East Nepal, reveals that melt is peaking in July and August, rather than becoming negligible in August, as suggested by Figure 21.

This difference in timing of the snowmelt season (February-August vs. May-November) may partly be due to the fact that the Khumbu study reflects meltwater from a glacier rather than from a seasonal snowpack at lesser elevations. A considerable portion of the flow now assigned to "interflow" in Figure 21 may thus in reality be due to meltwater contributions.

Also, it is difficult to envisage how hillside vegetation would survive a four-to-five month long dry period after "the interflow storage is exhausted in December and only baseflow accounts for runoff (Chyurlia, 1984).

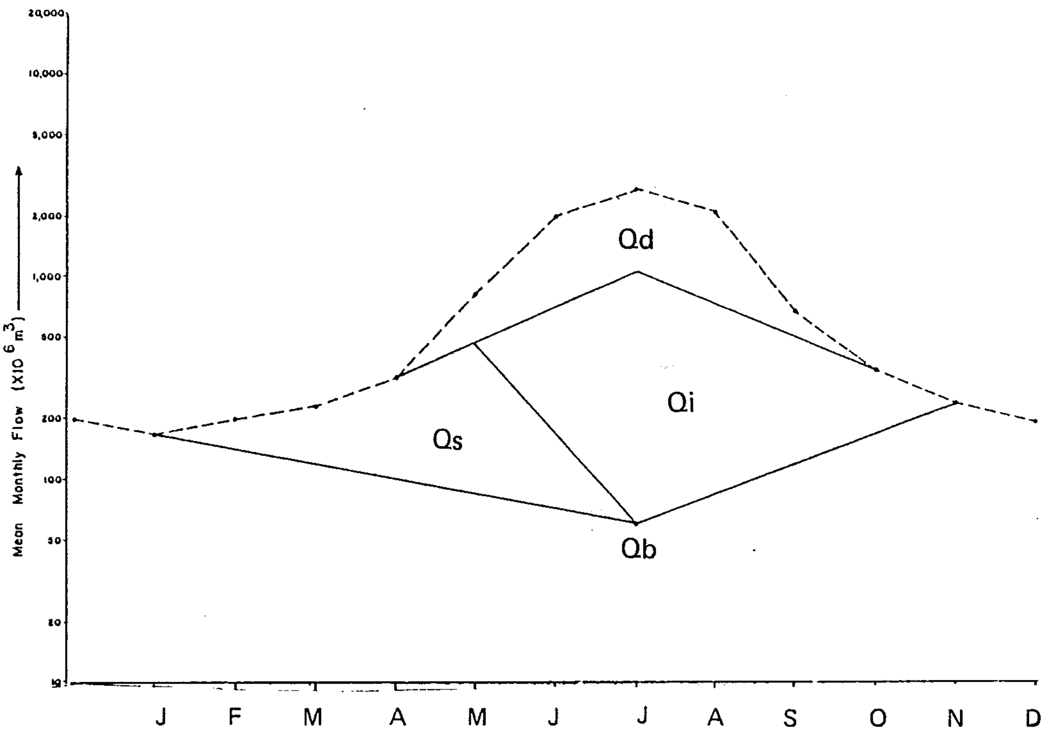


Figure 21 Hydrograph separation for the Seti river, West Nepal (after Chyurlia, 1984).

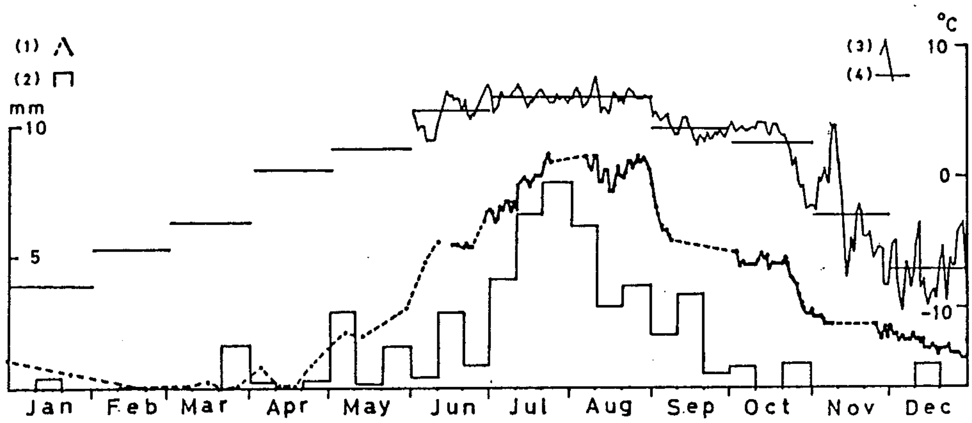


Figure 22 Water discharge (1) of the Imja Khola at Dingboche, precipitation (2) and air temperature (3) and (4) at Lhajung (eastern Nepal), 1974/75 (after Higuchi et al., 1976).

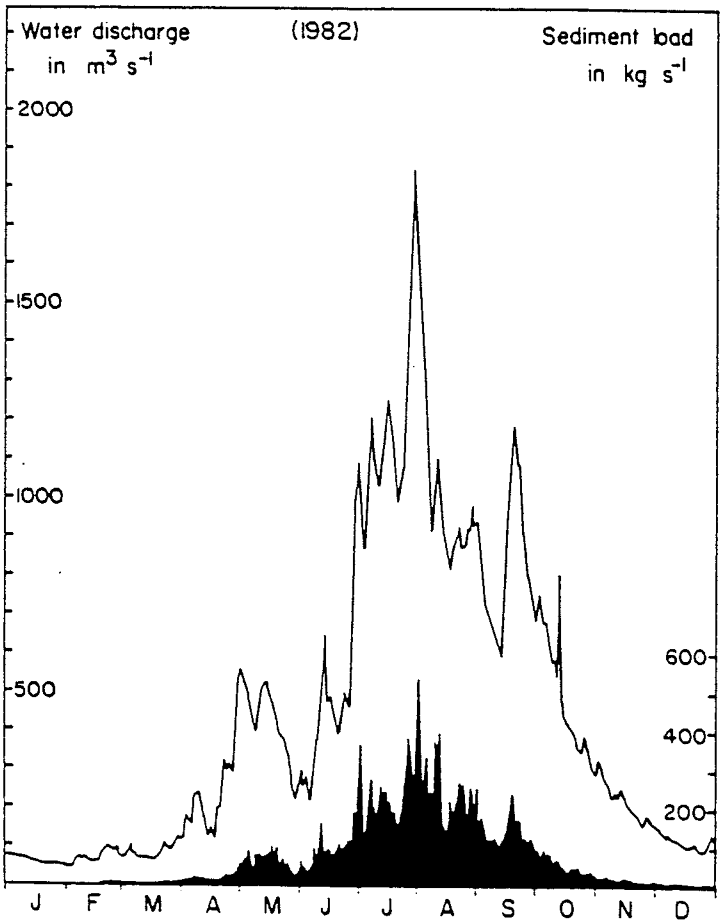


Figure 23 **Water discharge (m^3/sec) of the Burhi Dihing river, Assam, in 1982 (after Sarma, 1986).**

Rather, the saturated zones around (smaller) streams in the valleys, which supply the baseflow, should be seen as being fed by continuous (although steadily diminishing after the monsoon) "interflow" from the hillslopes (Hewlett & Hibbert, 1963; Figure 26).

The subject of flood-generating "direct flow" will be discussed in more detail in the next Section.

Whilst the Himalayan rivers, whose regimes were illustrated in Figure 20, carry 69% (Nyang He) to 83% (Tinau) of their annual flow during the summer monsoon months (June through September), this proportion is even higher for the southern rivers.

The Yamuna (346,000 km^2), for example, discharges 91% of its annual streamflow total into the Ganges during the monsoon. The rest is more or less equally distributed over the winter months (October-February) and the pre-monsoonal period (March-May).

Even more extreme in this respect are the Sind (93% flowing during the monsoon), the Chambal (95%) and the Betwa (97%) rivers (Jha et al., 1988).

A less extreme picture emerges for the Burhi-Dihing in Assam (Figure 23). Here, the contribution to total flow by streamflow in the spring months is comparatively greater as a result of an earlier start of the rains (Sarma, 1986; Figure 7).

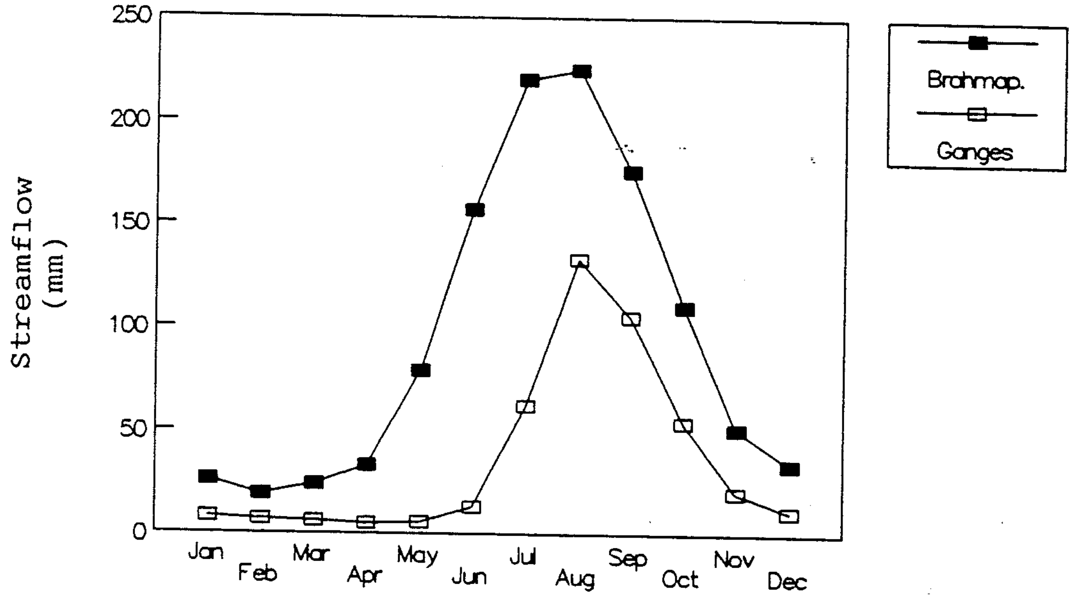


Figure 24 **Average monthly streamflow totals (mm) for the Brahmaputra at Bahadurabad Ghat and the Ganges at Farakka (based on data presented by Haroun er Rashid (1977) and UNESCO (1971) respectively).**

Naturally, all this has a bearing on the flow regimes of the Ganges and Brahmaputra rivers themselves (Figure 24). The Brahmaputra is seen to carry much more water per area unit than the Ganges (1155 vs. 430 mm/yr), reflecting the wetter conditions prevailing in the eastern part of the combined river basin (Figures 8 and 9).

Furthermore, the Ganges exhibits minimum flow rates in May, just before the onset of the monsoon, and peaks in August and September.

Conversely, the Brahmaputra has a minimum in February and starts to rise in March in response to the arrival of the pre-monsoon rains, culminating in July and August (Figure 24).

As such, August has always been the month in which widespread flooding in Bangladesh is most likely to occur (Haroun-er-Rashid, 1977). According to the same author (referring to the situation prior to the building of the barrage at Farakka), floods in the delta occurring between May and July are usually due to the Brahmaputra-Jamuna itself, and between August and October due to the combined flow with the Ganges.

The issue of peakflows will be considered further in the next section on runoff response to rainfall.

III.4 HILLSLOPE HYDROLOGICAL RESPONSE TO RAINFALL

When examining how peakflow rates in the Ganges-Brahmaputra River basin are generated, it is helpful to distinguish three scales, viz. a macro-, a meso- and a micro-scale (Ives et al., 1987). At the macro-scale one looks at the interaction between Highland and Lowland systems, e.g. the Himalayan block vs. the Indo-Gangetic depression.

At the intermediate scale a major tributary river basin, e.g. the Gandaki or the Tista, is considered. Such meso-scale catchments usually represent an array of environmental conditions, ranging from the alpine heights of the Great Himalaya to the tropical hills of the Siwaliks.

Micro-scale watersheds are small enough to fall entirely within a single specific zone, e.g. the Siwaliks, the Middle Hills, or the arid Trans-Himalaya. As such, they lack significant biotic zonation (cf. Figure 17) and are relatively homogeneous. Obviously, each meso-scale basin consists of a large number of micro-scale watersheds, and in turn the macro-system is made up of a number of meso-scale basins.

Similarly, the timing and magnitude of streamflow of the macro-rivers will reflect the flow distribution between the meso-scale watersheds, just like the response of the latter reflects the aggregate response of the micro-scale catchments of which they are comprised.

Rain falling on a hillslope may reach the adjacent stream channel in several ways (Figure 25a). If rainfall intensities exceed the capacity of the top-soil to absorb all rainfall, then the surplus runs off as "infiltration excess" or "Hortonian" overland flow (Horton, 1933; flow path Q(o) in Figure 25a).

This type of flow is rarely observed under forested conditions, but may occur after disturbance and degradation of the soil, as well as in sparsely vegetated areas (Dunne, 1978).

The remainder infiltrates into the soil profile and, depending on vertical and lateral hydraulic conductivities, soil moisture patterns and slope steepness, may take one of several routes to the stream (Figure 25a).

In the (relatively rare) case of deep, permeable and uniform deposits, the water will tend to travel vertically downward to the zone of saturation and from there onwards will follow a curving path to the stream (flowpath Qg in Figure 25a).

More often, however, soil permeability decreases with depth (Gilmour et al., 1987). Part of the water then percolates vertically until it meets an obstruction, such as a clayey B-horizon or bedrock. It is then deflected laterally and usually referred to as "interflow" or "throughflow" (Kirkby & Chorley, 1967; flowpath Qt in Figure 25a).

The bulk of this "throughflow" generally travels rather slowly through the soil matrix, feeding saturated sections around the streams, thereby maintaining the baseflow of the stream (Hewlett & Hibbert, 1963; Figure 26).

These near-saturated riparian zones in a catchment may act as a major source of stormflow, with the mechanism that produces this "quickflow" reflecting the prevailing geomorphological setting (Ward, 1984).

In the case of the situation depicted in Figure 25, i.e. a concave-shaped slope bottom bordering a stream, quickflow is often generated through the formation of what has been called a "riparian groundwater ridge" (Ward, 1984). During rainstorms the groundwater table, already relatively close to the surface in these riparian zones, rises rapidly due to throughflow contributions from upslope and direct rainfall on the spot, and may reach the surface (Figure 25b). Any rain falling on the now saturated area is unable to infiltrate and runs off along the surface together with the emerging throughflow ("return flow") as "saturation overland flow" (Flowpath Qo(s) in Figure 25).

If the process continues long enough, the entire foot slope may become saturated with water (Figure 25c) and become a source area for storm runoff. Since in such situations a considerable part of the quickflow is due to saturation overland flow, stormflow patterns in the stream will immediately reflect differences in rainfall intensity as they occur (Dunne, 1978).

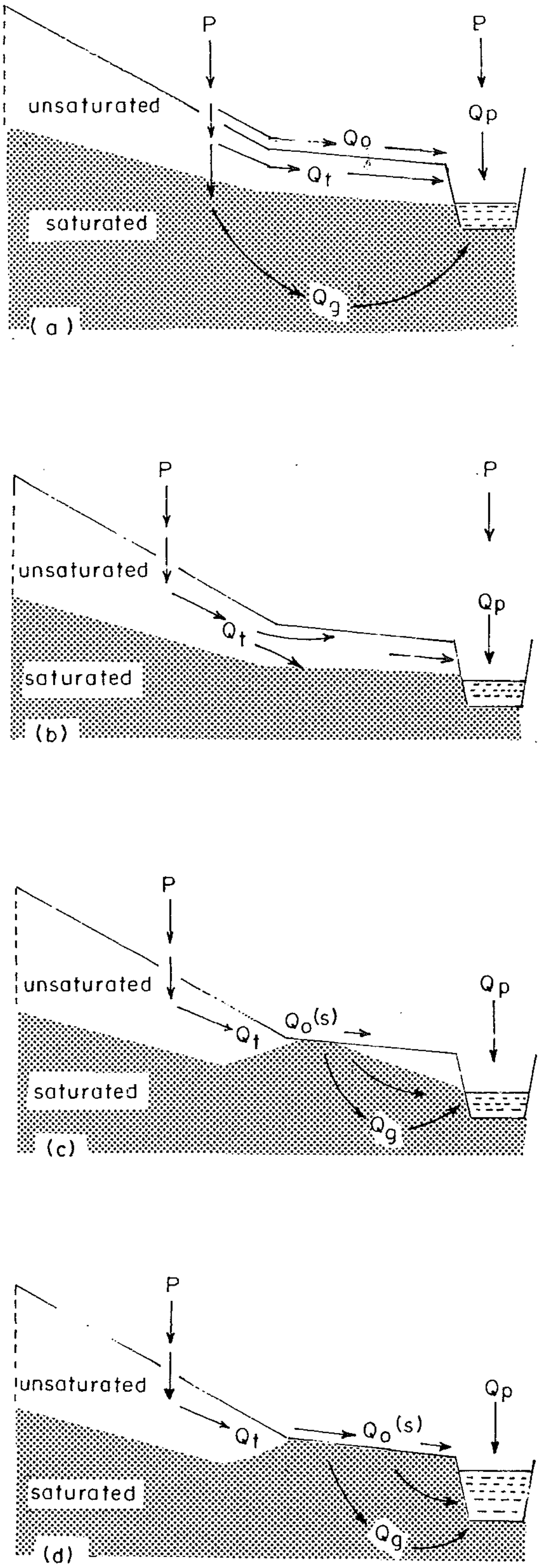


Figure 25.

(a) Flow paths of the sources of streamflow: Q_p is direct precipitation onto the water, Q_o is overland flow, Q_t is throughflow and Q_g is groundwater flow.

(b)-(d) The response of streamflow to precipitation in humid headwater areas: an integrated view (after Ward, 1984).

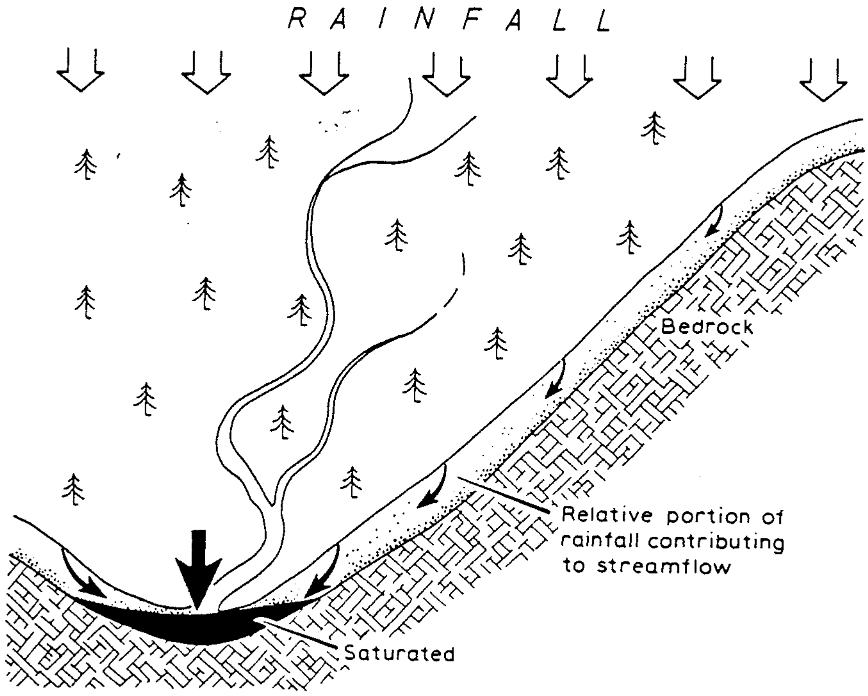


Figure 26 **The relative contributions of rainfall to streamflow (after Ward, 1984; based on an original diagram by J.D. Hewlett, 1961).**

However, where deep permeable soils overlie impermeable bedrock and where steep hillslopes border a narrow flood plain, there will be little scope for the generation of saturation overland flow, either in the valley bottoms or on the hillsides themselves.

To explain the immediacy of streamflow response in areas without appreciable overland flow of any type, it has been suggested that part of the throughflow travels through the upper soil horizons at a speed high enough to reach the stream channel during the storm (hence the frequently used term "subsurface stormflow").

Decayed root channels, animal burrows and other "macropores" have been advocated as conducts for such rapid throughflow (Whipkey, 1965; Mosley, 1982a). Whereas this may be true in certain cases, e.g. where subsurface "pipes" have developed (Jones, 1981), rates of water movement through soils are generally far too slow to enable "new" rainfall to reach the stream during a storm event (Dunne, 1978).

Therefore, Hewlett & Hibbert (1967) introduced the concept of "translatory flow", a "push-through" mechanism whereby each new addition of rainfall to a hillside displaces an approximately equivalent amount of "old" water, thus causing the oldest water to exit from the bottom end of the system and into the stream (cf. Figure 26).

Recent work by Pearce et al. (1986) and Sklash et al. (1986) has shown the overriding inportance of the "push through" mechanism in areas of steep slopes, permeable soils and high rainfall.

Naturally, an equivalent displacement of stored soil water by new rainfall can only be expected if the moisture storage capacity of the hillslope soil mantle is already filled or near full.

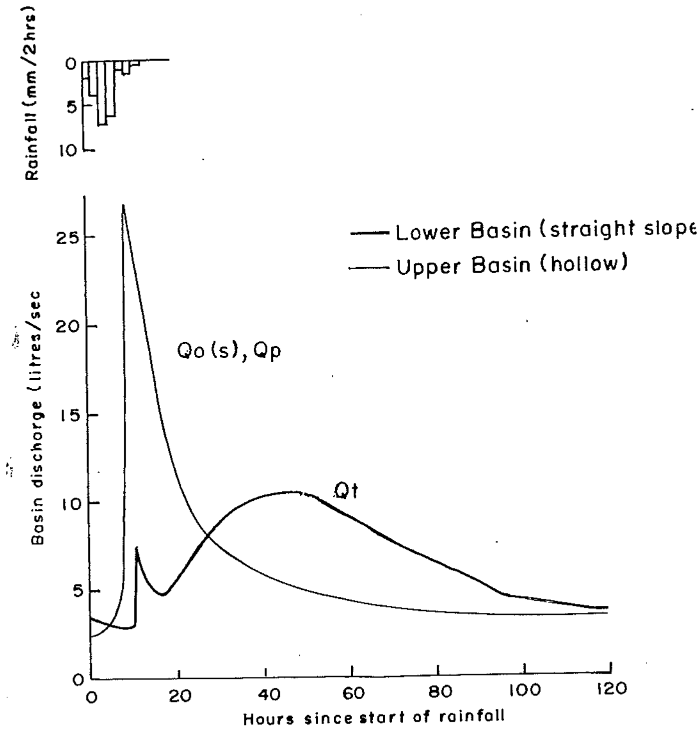


Figure 27 Storm-runoff hydrographs from two areas with contrasting topography within the East-Twin basin (0.2 km^2), United Kingdom (after Calver et al., 1972).

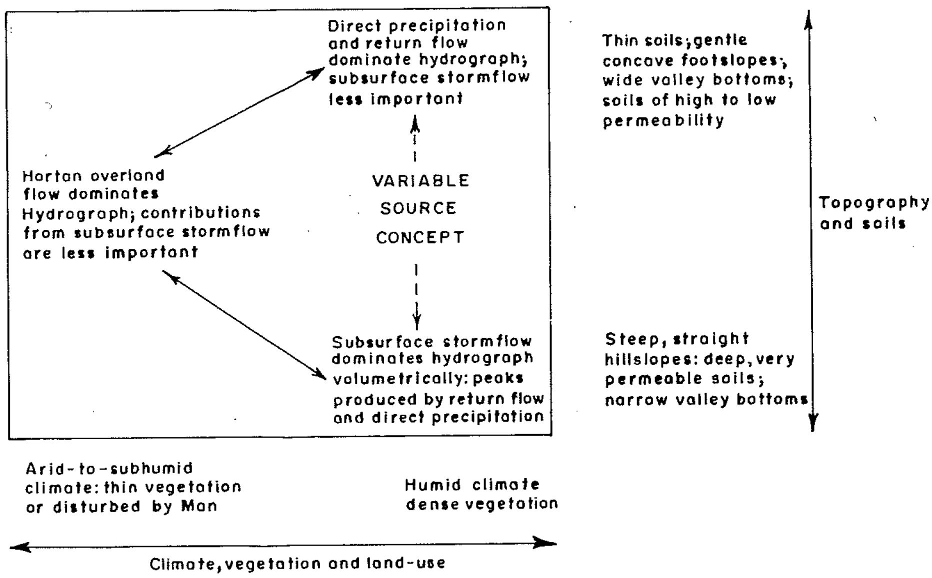


Figure 28 Schematic representation of the occurrence of various runoff processes in relation to their major controls (after Dunne, 1978). "Direct precipitation and return flow" equivalent to saturation overland flow.

In drier conditions rainfall inputs and/or displacements will be used to "top-up" the soil-moisture store rather than to maintain the chain of displacements. This implies that the mechanism will be effective most frequently after a period of rain and/or on the lower and moister parts of the slopes (Ward, 1984). Depending on the depth and initial moisture status of the soil and the magnitude of the storm, a peak in streamflow may occur shortly after the storm or up to several days later (Hewlett & Nutter, 1970; Sklash et al., 1986).

As shown in Figure 27, peaks produced by some form of overland flow tend to be much more pronounced than those generated by sub-surface flow types.

It is especially this shift from subsurface flow- to Horton overland flow-dominated storm runoff, that often accompanies certain changes in land use and produces an array of problems, as will be discussed in the next chapter.

Figure 28 summarizes the occurrence of the various runoff generating processes in relation to their major controls. As emphasized by Dunne (1978), the various modes of storm runoff should be seen as complementary rather than contradictory.

Plate 13

Rill and gully erosion on slopes underlain by shales in Dolpo, West Nepal (photograph by P. Laban).

How does all this translate to the Ganges-Brahmaputra region, and to the Himalaya in particular?

Leaving disturbed situations for the moment aside, one can make a general distinction between the dry Trans-Himalayan zone and the more humid parts of the mountains.

In the former, infiltration excess overland flow will dominate runoff peaks (Plate 13), whereas in the latter sub-surface stormflow (translatory flow) will be the major component of storm runoff (Figure 28; Pandey et al., 1983/4).

In view of the often very long and steep slopes and the relatively narrow valley bottoms (Plate 14), the contributions made by saturation overland flow from the riparian zone will be modest at best, with the exception of some poorly drained parts of the terai and perhaps the Dun valleys.

That (any type of) overland flow is indeed an extremely rare phenomenon on well-vegetated hillsides in the Middle Himalaya, was demonstrated convincingly by the work of Pandey et al. (1984) and Gilmour et al. (1987).

The former measured very low amounts of overland flow during two consecutive rainy seasons in a range of forest types in the Kumaon Himalayas, whilst the latter predicted low frequencies of occurrence of overland flow on the basis of rainfall intensity and soil hydraulic conductivity data for several sites in Central Nepal.

Interestingly, both groups of investigators observed relatively low rainfall intensities for most of the time (Table 3). In addition, the majority of storms was reported to be of relatively small magnitude (< 20 mm). Therefore, soil infiltration characteristics must have degraded considerably before appreciable quantities of overland flow could occur regularly under the rainfall regime prevailing in the Middle Himalaya (Pandey et al., 1983/4; Gilmour et al., 1987).

This aspect will be discussed more fully in Chapter IV on the role of vegetation and land-use.

Rainfall intensities are generally higher at lower elevations (Chyurlia, 1984) and as such the frequency of overland flow may be expected to increase for such areas as the Siwaliks.

Figure 29 illustrates the average intensity-duration curves for rainfall in Kathmandu (Middle Hills) and Dehradun (Indian foothills zone). Rainfall intensities are consistently higher at Dehradun, also for return intervals longer than a year (not shown).

Near-saturated top-soil permeabilities (Ksat) for forest soils from the two areas (Gilmour et al., 1987; Patnaik & Virdi, 1962) have been added for comparison (Figure 29). Based on these limited data, chances for the occurrence of overland flow would seem higher in the Indian Dun and Siwalik zone than in the Middle Hills of Nepal.

Grazing is a common practice in these forests and the herb/grass layer is often absent (Sastry & Narayana, 1984). According to Patnaik & Virdi (1962), top-soil infiltration capacities of these poorly structured soils were easily reduced by 50 % in case of compaction by grazing animals. Yet the value quoted in Figure 29 pertained to forest "with good leaf litter". Narayana & Sastry (1983) reported even lower values for grazed forests in the valley-bottom part of Doon valley, where soils are somewhat heavier. Carson (personal communication) observed widespread overland flow on Siwalik hillsides in Nepal during the monsoon.

Subba Rao et al. (1985) presented storm hydrographs for a densely forested (but grazed) headwater catchment in the Siwalik hills, which definitely suggest stormflow to be dominated by some sort of overland flow (Figure 30; cf. Figure 27).

These investigators also reported, that runoff would not be produced before the area had become thoroughly wetted up thoroughly, whilst flows would stop soon after the withdrawal of the monsoon in September.

In the absence of information on sub-soil permeabilities and groundwater levels in these catchments, it

Plate 14 Tributary valley of the Marsyangdi river near Tal, West-Central Nepal, with steep slopes and narrow valley bottom. Note the enormous boulders in the centre.

(a) Annual number of days with certain 5-minute rainfall intensities as recorded at Kathmandu, June - September, 1971-1979 (after Gilmour et al., 1987).
(b) Frequency distribution of rainfall intensity in different vegetation zones in the Kumaon Himalaya, June - September, 1981-1982 (after Pandey et al., 1984).

Rainfall intensity class (mm hr^{-1})	No. of days
0- 9.99	26.2
10- 19.99	9.8
20- 29.99	6.7
30- 39.99	4.9
40- 49.99	2.3
50- 59.99	0.8
60- 69.99	1.6
70- 79.99	0.7
80- 89.99	0.3
90- 99.99	0.6
100-109.99	0.1
110-119.99	0.0
120-129.99	0.4

Intensity classes (mm/ 30 min.)	Sites (% of events)		
	sal forest	pine forest	mixed-oak, Rianj-dominated forest
< 2	28.0	34.7	38.5
2–4	36.0	27.5	29.2
4.1–6	27.0	29.0	18.7
6.1–8	4.9	4.1	10.4
8.1–10	2.5	4.1	3.1
> 10	1.2	0	0
Total number of timed events	81	98	96

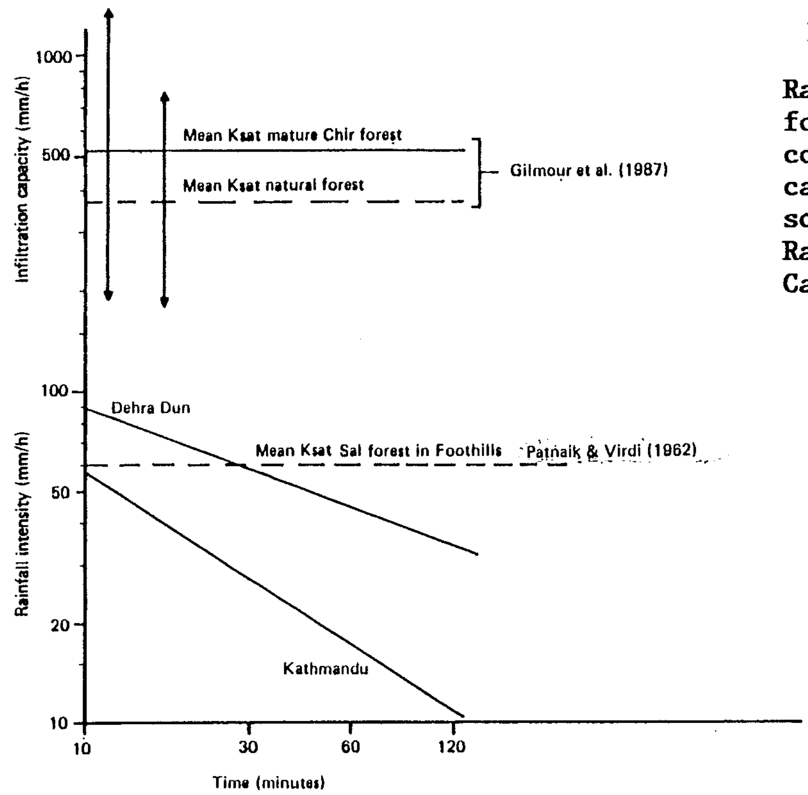


Figure 29.

Rainfall-intensity-durations for Dehradun and Kathmandu as compared with infiltration capacities of local forest soils. Rainfall data after Ram Babu et al., (1978) and Caine & Mool (1982).

is difficult to decide whether the overland flow involved is of the saturation or infiltrationexcess type. More detailed process-oriented work is needed to shed more light on these matters.

Other investigations of the hydrological response to rainfall of small catchment areas in the Himalayas are virtually limited to the studies carried out by the Central Soil & Water Research & Training Institute, Dehradun, in the valleys and foothills around Dehradun and Chandigarh (Narayana, 1987). Limited data have also been reported for the 1754-ha Bemunda catchment in the "Middle Hills" of Tehri Garhwal (Puri et al., 1982).

No information has been published for Nepal in this respect, although a catchment study has been initiated recently by the Topographic Survey Branch in the Dhulikhel area (P.B. Shah, personal communication; cf. Plate 4).

Average monsoonal runoff percentages (June-September 1976-1983) for two forested catchments underlain by deep and well-drained alluvial deposits in the Dun area ranged from 12% of incident rainfall (70-ha basin) to 19% (4.4 ha) according to data extracted from Annual Reports of CSWCRTI. Maximum (but still modest) runoff percentages were recorded for the month of August at the height of the monsoon (17 and 27% respectively).

The streams draining these flat basins are not perennial, suggesting that considerable amounts of infiltrated rainfall may contribute to deep drainage rather than to any type of lateral flow. In view of the very low relief of these basins (2-6%), however, it can not be excluded that valley-bottom saturation overland flow contributions to stormflow (Figure 25c) are significant as well.

These catchments are part of a larger experiment in which the hydrological effects of several conservation techniques are investigated (Ram Babu & Narayana, 1982; Sastry & Narayana, 1984).

Since the results of these experiments are often automatically assumed to be applicable to "the"

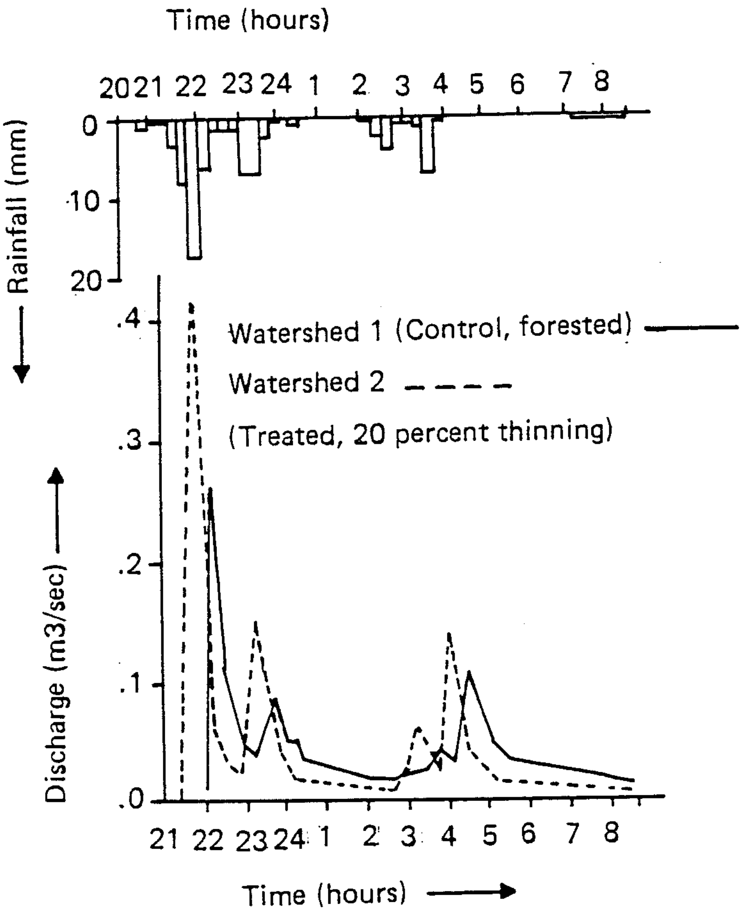


Figure 30.

Runoff response of two small forested headwater catchments in the Indian Siwaliks to 78 mm of rain (after Subba Rao et al., 1985).

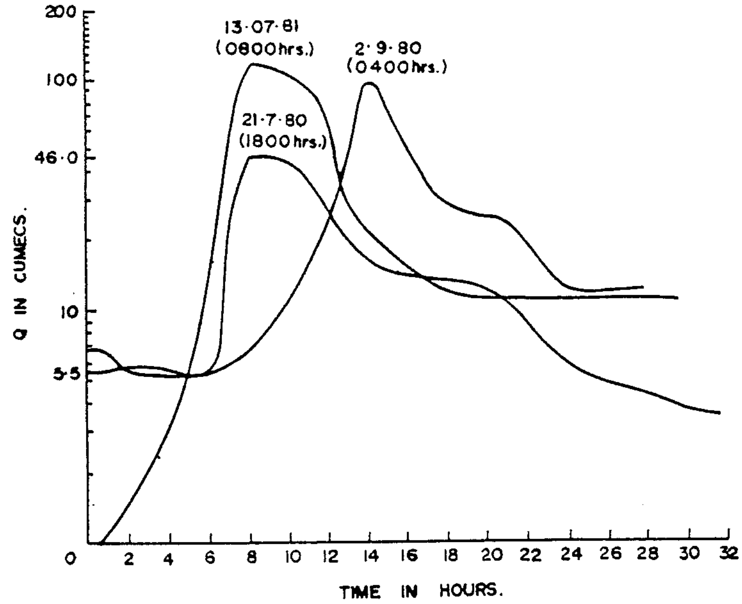


Figure 31.

Typical stormflow hydrographs for the pine-forested Bemunda catchment, Tehri Garhwal (after Puri et al., 1982).

Plate 15 **Actively eroding Siwalik landscape south of Hetauda, Central Nepal (photograph by P. Laban).**

Himalaya (e.g. Ghosh & Subba Rao, 1979; Bahadur et al., 1980; Narayana, 1987), it is important to remember that these catchments are by no means representative for any part of the Himalaya other than Doon Valley and similar areas of low relief (e.g. the Chitwan valley, etc.).

Limited data are also available for a very small (1.3 ha) catchment in the Siwaliks near Chandigarh (Gupta et al., 1974, 1975). In this somewhat steeper (25 %) area, which was covered with scrub, grasses and scattered trees, stormflow totals amounted to only 6 % of incident rainfall. Soils were deep and sandy, and presumably quite permeable (Plate 15).

As for the Middle Himalaya, it would seem as though the forested Bemunda catchment could be more or less representative of the overall hydrological conditions prevailing in the steeper parts of this zone.

According to Puri et al. (1982), streamflow rates during the rainiest months constituted 70-80% of incident rainfall, and on average 44% during the non-monsoonal months. Such high runoff coefficients for the rainy season could be seen as support for the importance of translatory flow during these months. Figure 31 presents a few storm hydrographs from Bemunda, whose recession limbs indeed suggest important subsurface contributions to stormflow (cf. Figure 27). In the absence of the corresponding rainfall patterns, this interpretation remains untested however.

In general, there is a need for more comprehensive catchment studies in the region, linking streamflow behaviour to hillslope hydrological processes in a quantitative manner. Such studies should preferably combine hydrometric and tracer techniques (Bruijnzeel, 1983; Sklash et al. 1986).

III.5 FLOODS: REGIONAL PATTERNS

In the absence of in-depth studies, several attempts have been made to predict catchment response (peakflow rate, stormflow volume) to rainfall in the (Lesser) Himalaya using empirical methods such as the rational formula (Meijerink, 1974; Chyurlia, 1984). The latter is essentially based on the concept that storm runoff is generated by infiltration excess overland flow (cf. Figure 25a). The formula consists of the product of catchment area, rainfall intensity and a "runoff coefficient" reflecting the lumped effects of soil and topographic characteristics, vegetation and anything not accounted for (Chow, 1964).

It will be clear from the above discussion of hillslope hydrological patterns that reality is more complex than envisaged by Horton (1933), certainly in the Himalayas. As such the results obtained with the rational formula should be treated with extreme caution.

Meijerink (1974) claimed to have obtained good results with the method for catchments up to 20 km^2 in the homogeneous Siwalik area around Dehradun. However, his method for assigning numerical values to the runoff coefficients associated with various types of soils, topography and land-use remained rather obscure. Also, when catchments situated outside the Siwaliks were included, results were less satisfactory, possibly because of variations in rainfall intensities. Nevertheless the calibration of Meijerink's "photo-hydrological" interpretation with streamflow results from small- to intermediate sized catchments seems a promising route to hydrological regionalization, provided that more reliable data on the spatial distribution of rainfall become available for the Himalayas.

An attempt at regionalizing storm runoff amounts from very small catchments (< 100 ha) in the foothill- and Siwalik area around Dehradun was also undertaken by Ram Babu & Narayana (1982).

Using long-term rainfall-runoff records for five catchments from Doon valley, ranging in size between 4 and 83 hectares, they calibrated a multiple regression model having as variables catchment area, channel length, storm duration and magnitude, maximum 30-minute rainfall intensity and soil wetness. The resulting coefficient of determination (R^2) was a modest 0.52, perhaps reflecting the fact that differences in landuse types had not explicitly been taken into account. Also, application of the model by the present writers to a small catchment in the Siwalik zone near Chandigarh (Gupta et al., 1974) produced a serious overestimation. This suggests that this "regional" model should not be used outside the context of the flat Doon valley, even though it was developed to predict stormflow characteristics for the region at large (Ram Babu & Narayana, 1982).

Chyurlia (1984) encountered similar difficulties when analyzing streamflow behaviour of larger (> 1000 km^2) catchments in Nepal. Although there were clear differences in amounts of "direct runoff" (cf. Figure 21) produced by the various basins, it was not possible for him to interpret these directly in terms of geological and rainfall factors.

Chyurlia circumvented the problem by examining the seasonal variation in runoff coefficients, which are largely determined by rainfall patterns. Statistically significant relationships were found to exist between monthly runoff coefficients and amounts of rain in the preceding month for each basin (Figure 32).

As expected, the highest runoff coefficients were observed for the months of August and September, when soils have been wetted thoroughly by the monsoon rains, thereby diminishing storage opportunities.

Interestingly, the constant and the coefficient of the regression equations appeared to correlate significantly with the percentages of catchment area occupied by the Middle and High Mountains physiographic regions (cf. Figure 4), at least for the 42.890 km^2 Karnali basin. Equally interesting was Chyurlia's finding that the maximum observed direct runoff coefficients only correlated significantly with percentage area in the Middle Mountains region (full data set).

In other words, the Middle Mountains physiographic region seems to have response characteristics to rainfall favouring greater quantities of direct runoff than do other physiographic regions, notably the High Mountains and, to a lesser extent, the Siwaliks (Chyurlia, 1984).

Whether this reflects higher rainfall totals or reduced infiltration characteristics is difficult to tell, since the analysis concerned river basins of considerable size (1000-57,000 km^2), with wide variations in geology and rainfall (Chapter II).

Chyurlia himself concluded from a principal component analysis of the entire data set that in the absence of more detailed catchment studies the total annual rainfall and percentage catchment area in the Middle Mountains were equally important and apparently inseparable.

Summarizing, our (quantitative) knowledge of storm-runoff generation in small and intermediate-sized catchments in the Himalaya is limited.

As pointed out by Chyurlia (1984), the study of spatial variations in runoff coefficients and their dependence on rainfall and other factors, constitutes an area of research that should be given high priority.

The role of vegetation and land use in this respect will be discussed in Section IV.1.3 and should certainly not be underestimated at this scale.

The timing and distribution of rainfall becomes even more important with respect to the occurrence of peakflows at the meso- and macro-scale levels than in smaller catchments.

Clearly, stormflow generated by heavy rainfall in one part of a large basin will be "diluted" by lower flows from other parts of the basin receiving less or no rainfall at that time. Therefore, although the ***average*** seasonal flow patterns for the major Himalayan rivers are quite similar (Sharma, 1977), the timing of ***actual*** peak discharges on one river does not necessarily coincide with that on other tributaries.

Alford (1988b) conducted a preliminary peakflow analysis for the Ganges at Farakka and all major Nepalese tributaries as recorded between 1967 and 1976. He was unable to detect any correlation between the occurrence of maximum flows at Farakka and those for the Himalayan rivers. More importantly, there was also no correlation in this respect between the Himalayan rivers themselves. Alford (1988b) concluded that "there are no uniform 'wet' and 'dry' years dominating the entire region", or even Nepal. Although his analysis was preliminary and covered only one decade, it reached essentially the same conclusion as that of Mooley & Parthasarathy (1983), who studied regional rainfall records over the period 1871-1980 (see Section II.3.2).

In other words, spatial variations in rainfall patterns seem to override any other factor in the generation of floods at this scale, both directly and indirectly (see below). As such, extreme streamflow events must reflect the occurrence of too much rain falling persistently over a large area.

As shown earlier in Figure 13, large parts of the Ganges-Brahmaputra River basin may receive substantial amounts of rain in a single day. Figure 33 shows an example from Assam, where an area of almost 4000 km^2 received at least 300 mm of rain in two days, with the central 500 km^2 of the storm receiving more than 600 mm. The total area enclosed by the 100 mm isohyet amounted to more than 22.000 km^2 (Raghavendra, 1982). Similar examples, both for the upper Brahmaputra and Ganges areas, can be found in Pal & Bagchi (1975), Pant et al. (1970), and in Dhar et al. (1982b).

It will be clear from the above that the source of flood generation is not restricted to extreme rainfall in

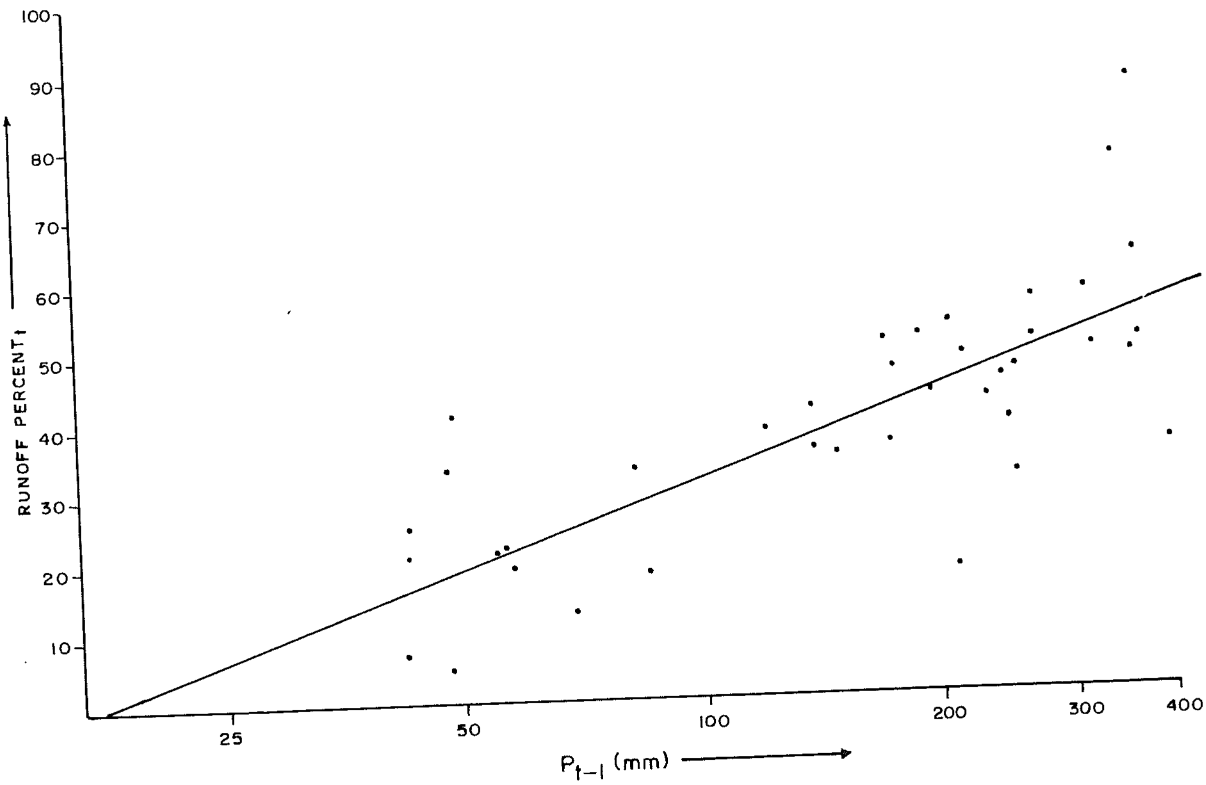


Figure 32 Semi-logarithmic relationship between percentage of direct runoff and antecedent monthly rainfall in Nepal (after Chyurlia, 1984).

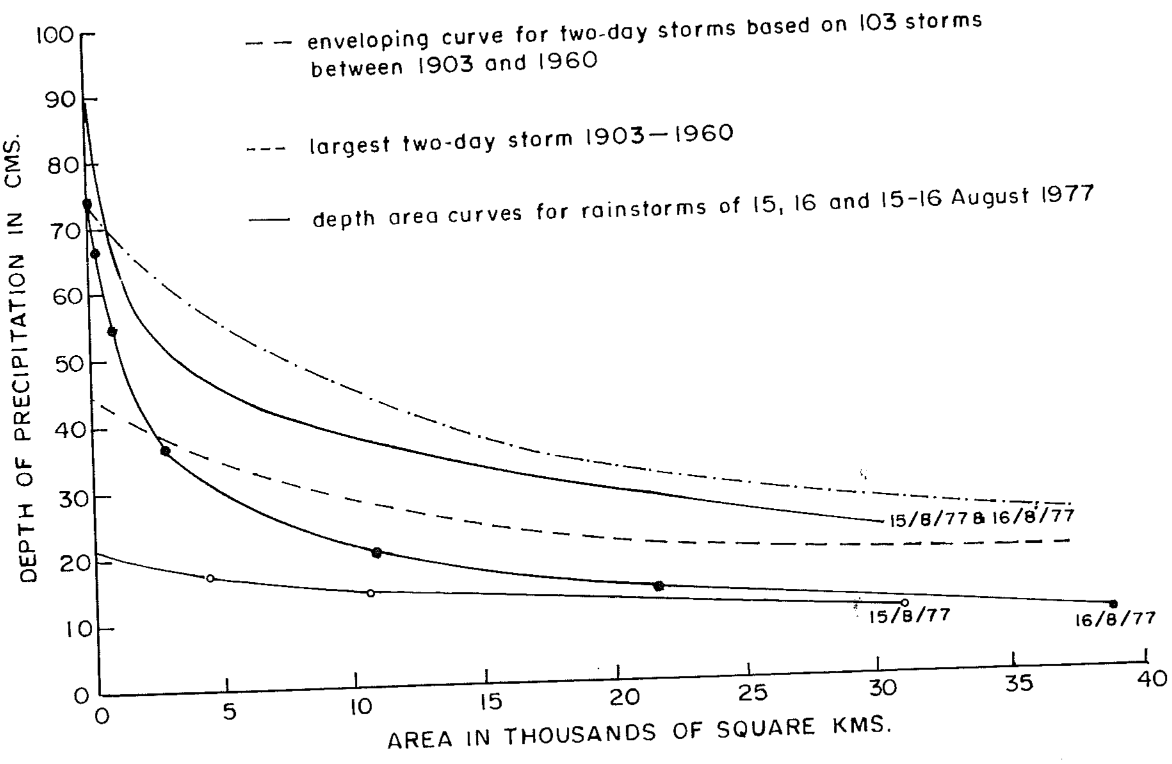


Figure 33 Depth-area curves for extreme rainfalls in the Assam valley (after Raghavendra, 1982).

previously wetted uplands. Excessive rain falling on the flood (!) plains themselves (cf. Figure 13) is capable of causing flooding overnight, especially since groundwater tables in the plains are close to the surface during the monsoon and storage opportunities limited (Rudra, 1979).

Indeed, Sen (1979) considered local overland flow and groundwater contributions extremely important in determining the height of flooding in the downstream part of the Gangetic plain.

Another important aspect of flooding in the plains relates to the so-called "backwater effect": the blocking at a confluence of the flow in one river when the other is in spate (or has a higher discharge). Such an artificial rise in water levels may extend for 100-150 km from the confluence and often leads to increased sedimentation in the affected river stretch (Figure 34; Anonymous, 1983).

The effect is illustrated in Figure 35 for a part of the Ganges near Patna, where water levels are raised by 6 to 8 metres due to the joining of the mighty Gandaki river (cf. Figure 1). The implications of backwater effects for the occurrence of flooding in the flat terrain of the plains need no further comment.

At a much larger scale the backwater effect is demonstrated by the joining of the Ganges and the Brahmaputra (Figure 1). Although the flows on the Ganges are much smaller than those on the Brahmaputra, when expressed as a layer of water over the respective total catchment areas (Figure 24), the actual volumes of water meeting in August and September are very similar (Figure 36; Plate 16). Also, there is the effect of rain water becoming "trapped" on land in the floodplain due to the high water levels in the main rivers.

A more indirect effect of extreme rainfall on flooding levels consists of the triggering of large-scale mass movement, especially when earth tremors occur during such wet spells (Goswami, 1985). These landslides introduce enormous amounts of sediment to the drainage system (Carson et al., 1986; see also section III.6).

The effect may be two-fold: (1) the extra sediment may cause aggradation of the river bed further downstream, thereby increasing the water level in general and flooding hazards in particular (Pal & Bagchi, 1975), and (2) the largest of these landslides may temporarily dam a river, producing a devastating surge of water and sediment after the debris barrier gives way. Such flood waves may attain heights of 15 to 20 m (Singh et al., 1974; Mahmood, 1987).

Increased sedimentation may also be the result of river training works or barrages. Rudra (1979) for example reported deposition of sediment upstream of the Farakka barrage in the Ganges, because of which the possibilities of flooding have been increased in various parts of Bihar.

Similarly, the relatively silt-free water released from the barrage tries to recover its lost load by eroding the river bed downstream of Farakka. This material is in turn deposited further downstream and hampers the navigability of the river and may even endanger the future of Calcutta Port (Rudra, 1979).

It would seem that the devastating floods which occurred in Bangladesh during the 1988 monsoon were the result of a combination of the above factors, although it will probably never be possible to separate them (Rogers, 1988).

Another and equally destructive flood-producing phenomenon is formed by glacial lake outbursts, which occur from time to time in the High Himalaya (Ives, 1986). These sudden bursts of lakes dammed by glacier ice or moraines may produce volumes of water that are often an order of magnitude greater than normal rain-derived peakflows and may travel tens of kilometres downstream, transporting colossal amounts of debris (Galay, 1987; Plate 17). Unfortunately, their

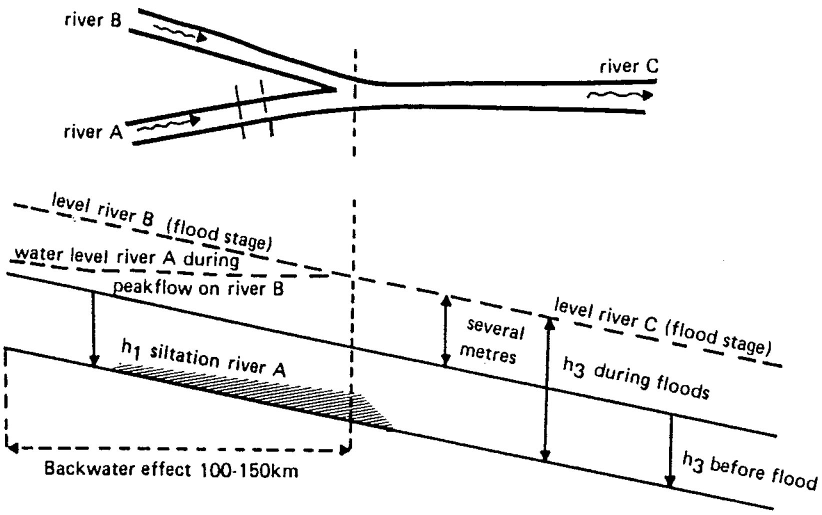


Figure 34 Schematic illustration of the "backwater effect" (modified from Dutch Inland Water Transport Mission, 1982-83).

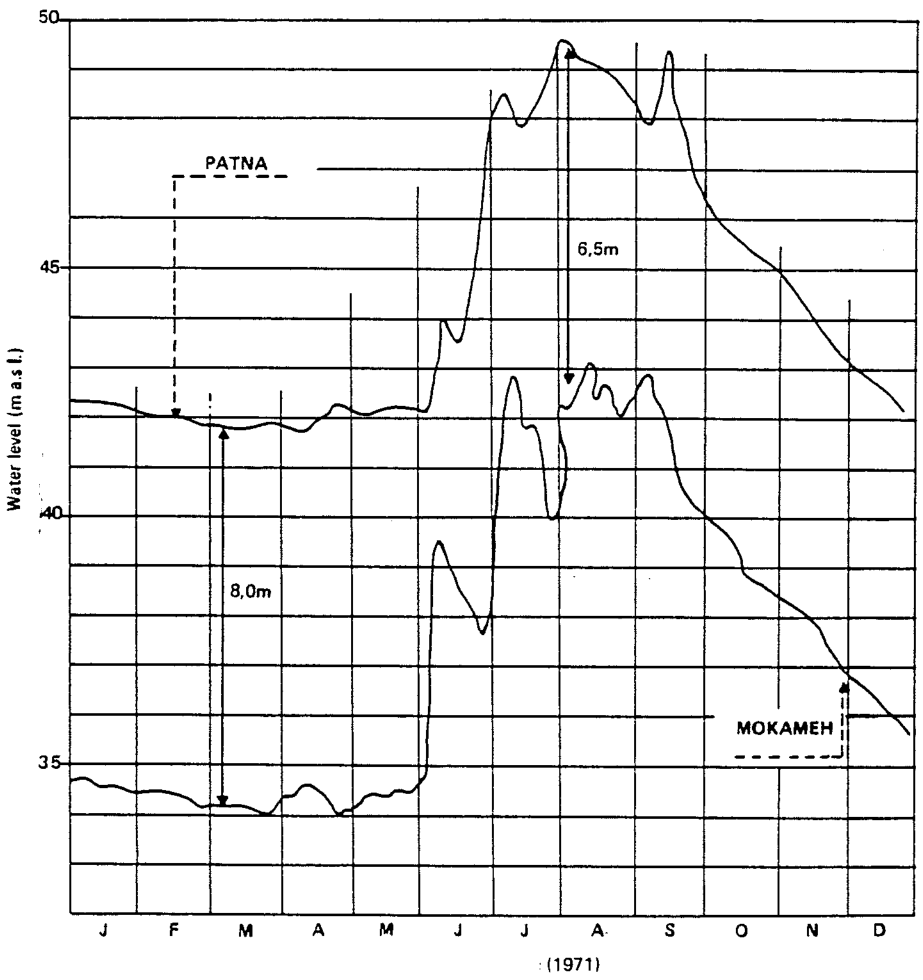


Figure 35 Example of the "back water effect" on the Ganges near Patna (after Dutch Inland Water Transport Mission, 1982-83).

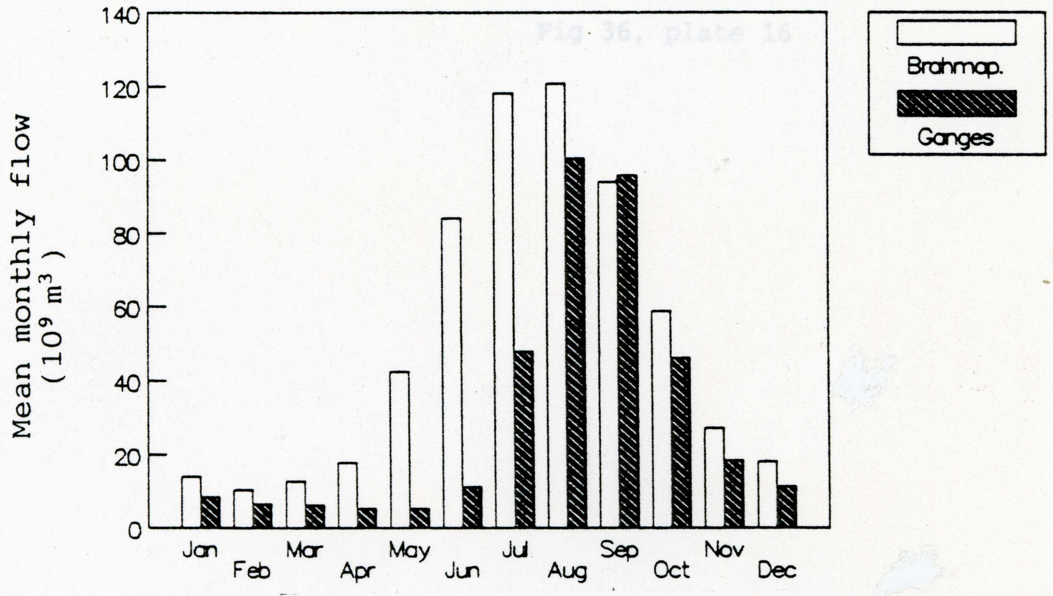


Figure 36. **Average monthly discharge of the Brahmaputra at Bahadurabad and the Ganges at Sara Bridge (based on data presented by Haroun er Rashid, 1977).**

Plate 16 **The confluence of the rivers Ganges (left) and Jamuna (right) in Bangladesh, June 1986 (photograph by G.J.Klaassen).**

Plate 17 A small glacial lake outburst flood disrupting a bridge on the Dudh Kosi river, Central Nepal, July 1984 (photograph by J. Desloges).

very nature makes "GLOFs" extremely difficult to predict (Ives, 1986).

All in all, there seem to be enough reasons why the lower Ganges-Brahmaputra plains should be liable to regular and extensive flooding, even when the neighbouring uplands were fully forested. The question to what extent recent deforestation in the mountains has exacerbated the flood problem will be addressed in Section IV.1.3.

III.6 RIVER SEDIMENT LOADS

In his reconnaissance survey of amounts of sediment transported in suspension (as opposed to rolling along the river bed) to the seas by the major rivers of the world, Holeman (1968) concluded that Asia's rivers were by far the greatest contributors, possibly supplying up to 80% of the world total.

Of the Asian rivers, the combined Ganges-Brahmaputra basin ranked as the first, with an estimated annual sediment yield of 2.4 billion tons (15 t/ha), of which the Ganges alone contributed about two thirds (Holeman, 1968).

A more modest estimate (1.67 billion tons, or 11.3 t/ha) was published more recently by Milliman & Meade (1983). Although sediment loads transported by such very large rivers can never be determined with great precision, it will be clear that the amounts of sediment carried by the two rivers must be enormous.

This is well illustrated by a comparison with the sediment loads for such tropical rivers as the Orinoco (0.9-2.1 t/ha/yr) or the Mekong (2.1 t/ha/yr) (Holeman, 1968; Milliman & Meade, 1983). Since both streamflow totals and basin size for these rivers are comparable to those of the Ganges, such differences in erosion must

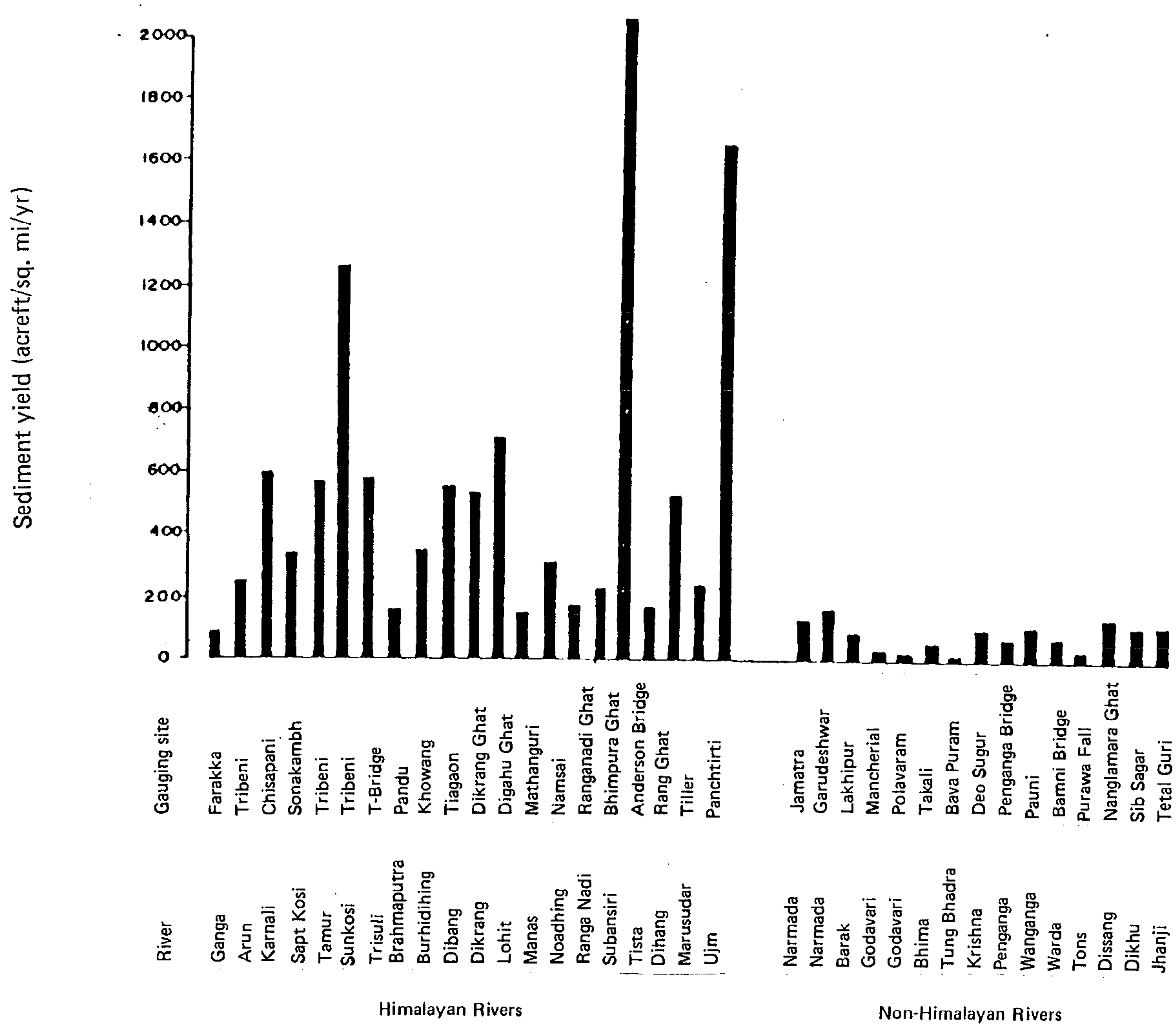


Figure 37 **Sediment yields for Himalayan and Peninsular rivers in India (after Gupta (1975) in Tejwani, 1985).**

reflect differences in basin relief and geology (Gregory & Walling, 1973).

As such, major contrasts in sediment transport may be expected between "Himalayan" and "Peninsular" rivers within the Ganges-Brahmaputra river basin (Figure 37).

A glance at Figure 37 not only reveals much lower sediment loads for non-Himalayan rivers, but also rather dramatic differences between the Himalayan streams themselves. In some cases such differences will reflect contrasts in basin size, since sediment yields (per km^2) generally tend to decrease with basin area; larger basins usually have more opportunities to store sediment, e.g. on their floodplains (Dunne, 1977).

Furthermore, as was observed for annual streamflow totals (Section III.1), the larger the portion of a catchment in the dry Trans-Himalaya the lower will be its sediment load (Das, 1968). Finally, certain rock types (e.g. phyllites, shales or unconsolidated sandstones) are much more susceptible to erosion and/or mass movements than others (see section IV.2), resulting in widely different stream sediment loads under otherwise similar conditions (Rawat, 1985; Biksham & Subramanian, 1988).

As such, each "meso-scale" basin naturally represents a more or less unique combination of the above variables. A good example of such local contrasts in sediment yield is exhibited by the three main branches of the Kosi above Tribeni in East Nepal (Table 4).

In the Tamur basin, the combination of high streamflow rates (reflecting high rainfall and meltwater contributions) and steep and unstable terrain with relatively few opportunities to store sediment, results in an impressive rate of sediment transport (cf. Figure 37). However, the influence of human activities in this area should not be underestimated (Brunsden et al. 1981). This aspect will be discussed more fully in Chapter IV.2.

Sediment yields in the Arun basin on the other hand are quite low, reflecting the fact that 90% of its area receives very little precipitation. In addition, a significant part of this dry headwater area consists of flat and broad valley bottoms which act as effective traps for sediment coming from the slopes.

However, if we take the 10 m^3/ha/yr quoted for the Arun in Table 4 at face value, the contributions to the overall sediment load made by the High and Middle Himalaya physiographic regions in the Nepalese part of the basin must be considerable.

Taking 1.5 t/ha/yr as a representative value for the amounts of sediment transported by South Tibetan rivers of this magnitude (Guan & Chen, 1981), it follows that the remaining 10% of the basin should supply about 120 t/ha/yr. Although the latter figure may seem excessive, it certainly points to the importance of the High and Middle Himalaya regions as contributors of stream sediment (Carson, 1985).

The Sun Kosi occupies an intermediate position between the Arun and Tamur, as could be expected on the basis of its annual streamflow total and percentage catchment area in the dry Himalaya (Table 4).

Table 5 explores a little further the spatial variations in suspended sediment loads between the various physiographic units (cf. Figure 3).

It would seem from these limited (semi-long term) data that the Middle Himalaya supplies the greatest quantities of (fine) sediment to the streams. This accords with the observation of Chyurlia (1984) that the zone also produces the largest amounts of stormflow (see section III.4).

Virtually no reliable data are available for streams in the High Himalaya and it may well be that suspended sediment loads in this highly glaciated and high-energy environment are larger than suggested in Table 5.

Alternatively, it could be argued that much of the sediment released by physical weathering in this zone is initially too coarse to be transported in suspension (Plates 2 and 3) and will move along the river bed until sufficiently broken for suspended transport (Galay, 1987). Further work on this matter is desirable.

As already indicated, rivers in the Trans-Himalayan zone and streams draining the old plateau in the south carry very little sediment (Table 5).

Not all of a river's sediment load is transported in suspension. A significant, but often unmeasured part moves along the channel as bedload. Although separate measurements of bedload transport in Himalayan rivers are rarely available (Galay, 1987), there is a growing body of information regarding total (i.e. suspended plus bed-) loads deposited behind dams erected on several major rivers in the area (Table 6).

Recorded rates of sedimentation in the Himalaya again vary widely. The low value found for the Bhakra reservoir must reflect the fact that a considerable part of the upper Sutlej catchment is situated in the dry Trans-Himalaya. In addition, the lower parts of the basin do not receive much rainfall either, due to their position in the far west (Figure 8). Furthermore, there may be a scale effect as well, the Bhakra basin being the largest Himalayan catchment quoted in Table 6.

The Kalagarh reservoir is fed by streams that mainly drain the Middle

Table 6. Rates of sedimentation for major reservoirs in and around the Ganges river basin

Name of reservoir	Siltation rate ($m^3/ha/yr$)	Period of measurement	Catchment area (km^2)
Himalaya			
Bhakra (Sutlej)[1]	6.9	1959-1965	56.875
	6.0	1965-1970	
	5.9	1970-1981	
Kalagarh (Ramganga)[2]	57.5	1975-1978	3.105
	52.5	1978-1981	
Southern Plateau			
Mayurakshi (Mayurala)[3]	16.5	1955+	1.790
Maithon (Baskar)[3]	14.3	1956+	5.205
Panchet (Damodar)[3]	10.8	1956+	9.815
Matatila (Betwa)[3]	4.4	1958+	20.750
Gandhisagar (Chambal)	9.6	1960+	21.875

[1]Singha & Gupta (1982); [2]Anonymous (1987); [3]Gupta (1983)

Himalayan and Siwalik zones. The high siltation rate is in agreement with other observations for this zone (Table 5).

As for the southern plateau, there may again be a tendency for larger basins to have somewhat lower siltation rates (i.e. per km^2). However, this may also be related to differences in rainfall as the Maithon and Panchet reservoir areas receive more rainfall than those situated further west (cf. Figure 8; Lal et al., 1977).

The very high amounts of sediment carried by most major Himalayan rivers cause them to adopt a braided pattern upon reaching the piedmont zone (Figure 6; Plates 7 and 9).

A good example is the Kosi river in eastern Nepal, which has not been able to cut a deep and stable bed for itself after leaving the Chatra gorge.

During times of high flow, the river easily overtops its shallow banks and spreads over a vast expanse that may in places be up to 30-40 km wide (Pal & Bagchi, 1975). Due to the continued deposition of sediment, the river bed has risen to several metres above the surrounding plain, thereby creating a highly dangerous situation (Zollinger, 1979).

However, as shown in Figure 38, the Kosi has been notoriously unstable for more than 250 years (Gole & Chitale, 1966; updated by Galay, 1987). Interestingly, the shift has been progressively westwards (Pal & Bagchi, 1975).

Also, according to Sharma (1977), the rivers lying east of Butwal (Nepal's Western Development Region) are shifting towards the west, and those west of Butwal towards the east, suggesting a tectonic origin for such movements. Morgan & McIntire (1959) have described several other cases of major river shifts in their analysis of the Quaternary geology of the Bengal basin (see also Coleman (1969) and Klaassen & Vermeer, 1988).

Tejwani (1985) discussed the work

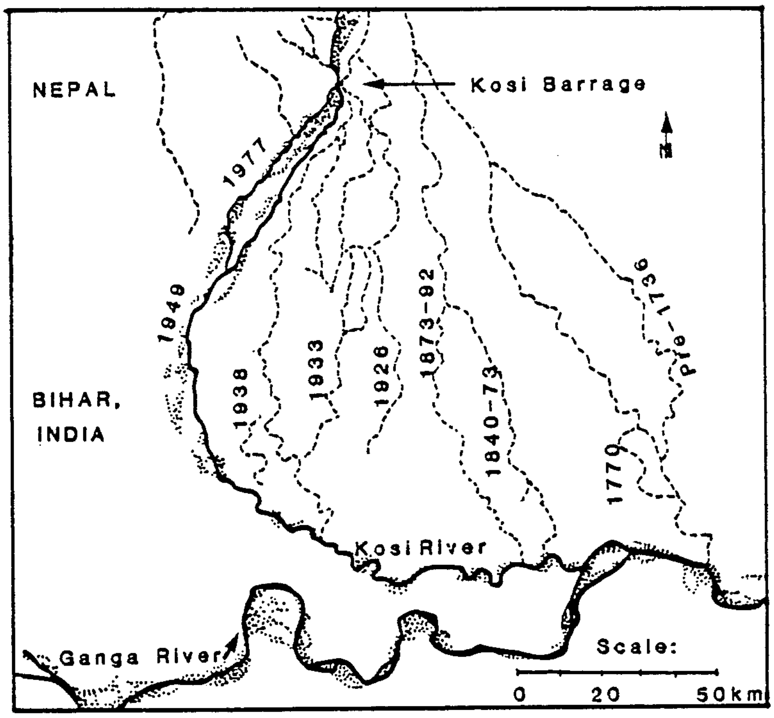


Figure 38 Shifting of the Kosi River over its own alluvial fan, 1736-1977 (after Gole & Chitale, 1966; updated by Galay, 1987).

of Gupta, who analysed sequential satellite imagery and observed substantial widening of river beds in the Terai zone between 1972 and 1979. The effect was ascribed to upland deforestation. As such, it is most important to monitor such changes in river morphology closely to reveal any further trends.

In view of the strongly seasonal character of most of the rivers in the region (Figure 20), the bulk of the annual sediment transport will take place during the summer monsoon. Also, major variations between years are to be expected as a result of inter-annual variations in rainfall distribution (Singha & Gupta, 1982; Figure 39).

It follows that long-term observations of sediment discharge are needed in order to arrive at meaningful estimates of average values. For example, suspended sediment transport in the Bhagirathi river (Tehri Garhwal, catchment area 7510 km^2) between 1973 and 1981 varied by a factor of 5.4 (Figure 39).

Therefore, the results of short-term studies, such as the comparative investigations of Subramanian (1979), Abbas & Subramanian (1984) and Jha et al. (1988), should be seen as representing an order of magnitude at best.

The inadequacy of a limited sampling program was demonstrated rather dramatically in a study of sediment transport on the Godavari, one of the rivers draining the southern plateau area (Biksham & Subramanian, 1988).

Based on occasional sampling during the various seasons over a period of three years, the authors computed a mean suspended sediment concentration

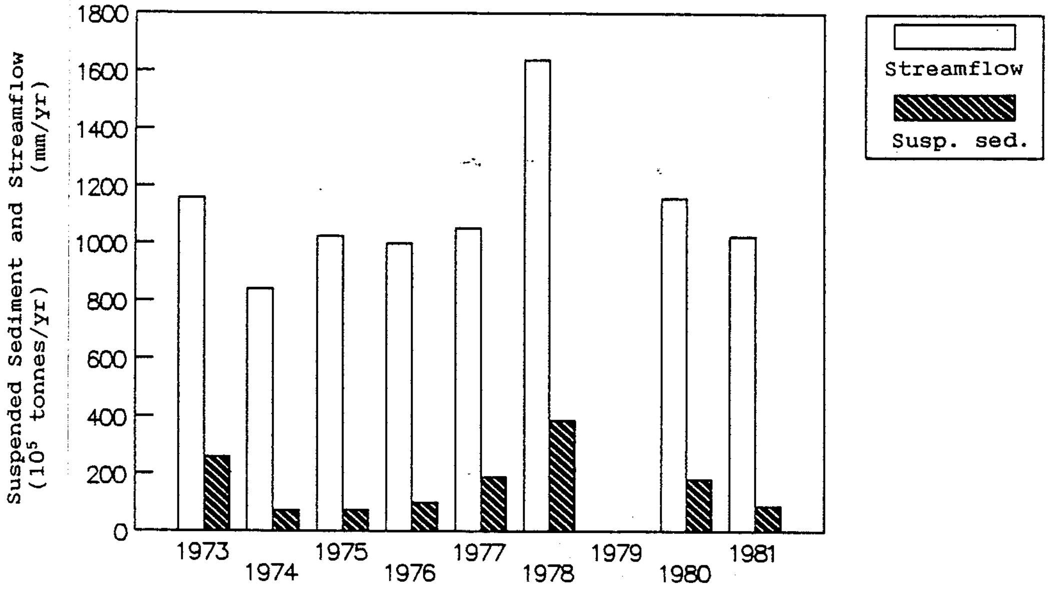


Figure 39 (a) Annual streamflow and suspended sediment load for the Bhagirathi river, Tehri Garhwal, 1973-1981.

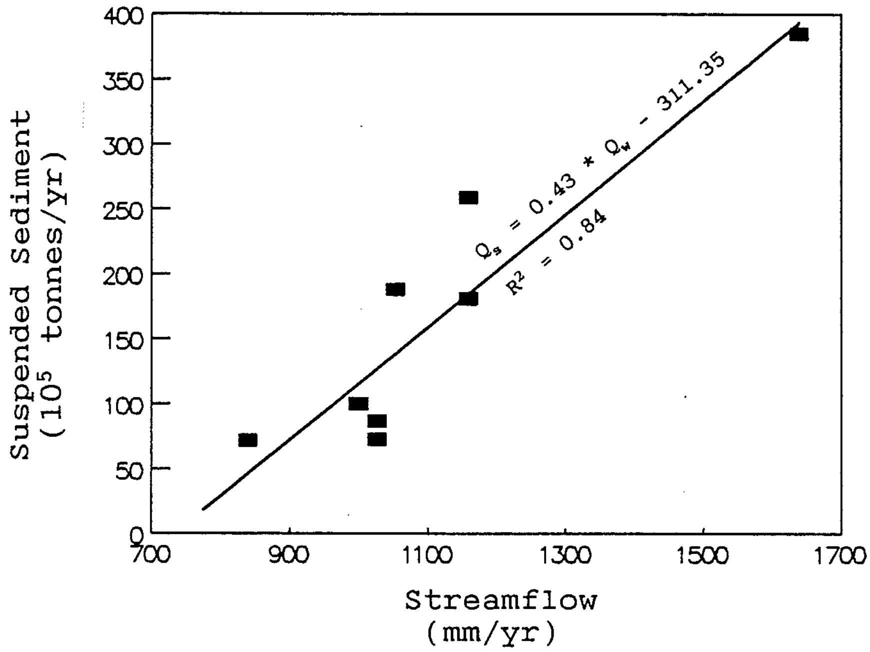


Figure 39 (b) Relationship between annual streamflow and amount of sediment carried in suspension by the Bhagirathi river (based on original data of Singha & Gupta, 1982).

of 770 mg/l. The corresponding value, based on a year's sampling on a daily basis by the Central Water Commission, amounted to 1525 mg/l, whilst the 10-year average concentration (daily observations) was 1845 mg/l (Biksham & Subramanian, 1988).

In this case, the annual suspended sediment load would have been underestimated by as much as 240%. The practical implications of the above, e.g. with respect to computations of the useful life of reservoirs, are of course profound.

Sampling strategies should obviously concentrate at an adequate coverage of the flows during the summer monsoon. In the already cited study of Singha & Gupta (1982) on the Bhagirathi river, as much as 92 to 99% of the annual suspended load was carried between June and September. Similarly high values have been reported for several rivers in Nepal (Sharma, 1977) and on the southern Plateau (Biksham & Subramaman, 1988), as well as for the Brahmaputra itself (Goswami, 1985).

In addition, within the rainy season, the contributions made by a few days of abnormally high flows can be disproportionally large. On the Godavari, Biksham & Subramarian (1988) observed values of 5 to 9% of the annual load being transported on individual days, whilst values of up to 64% could be contributed by the flow of a single month.

Similarly, Sarma (1986), working on the Burhi Dihing in Assam (Figure 23), reported that 50% of the total load was carried in less than 7% of the time (Figure 40).

Day-to-day variations in river sediment concentrations during the monsoon are considerable, with peak values being especially frequent in the second half of the rainy season (Figure 41b; Carson, 1985; Biksham & Subramanian, 1988).

Since surface erosion rates in the area are highest when vegetation cover is still sparse and rainfall intensities high, i.e. at the beginning of the rains in June (Figure 41a; Impat, 1981), some other mechanism of

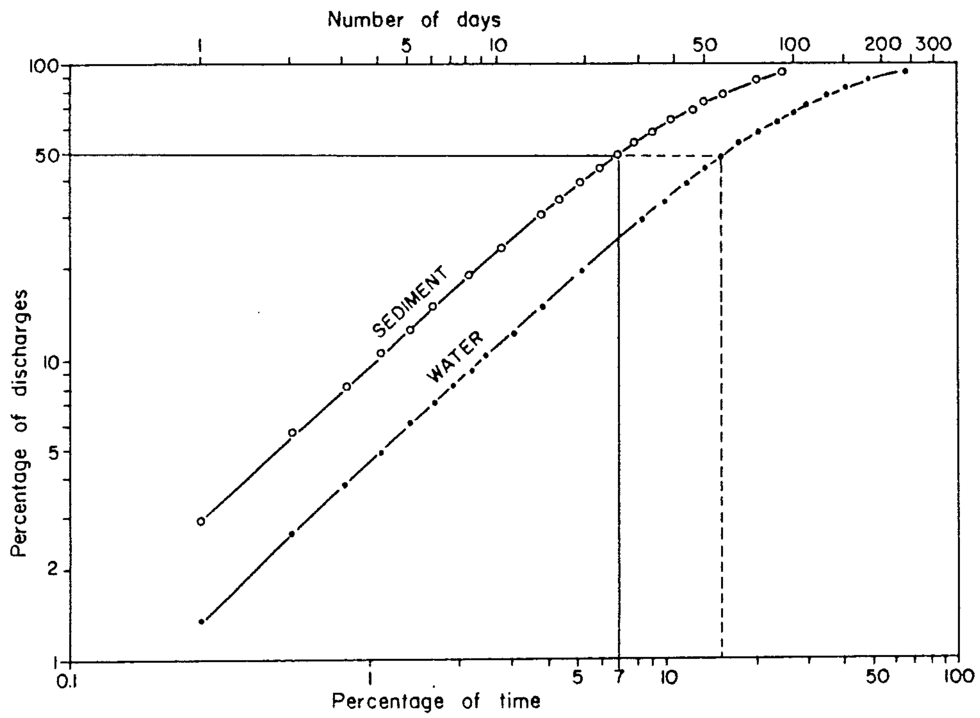


Figure 40 **Cumulative percentages of suspended sediment- and water discharges on the Burhi Dihing river against cumulative percentage of time for the year 1974 (after Sarma, 1986).**

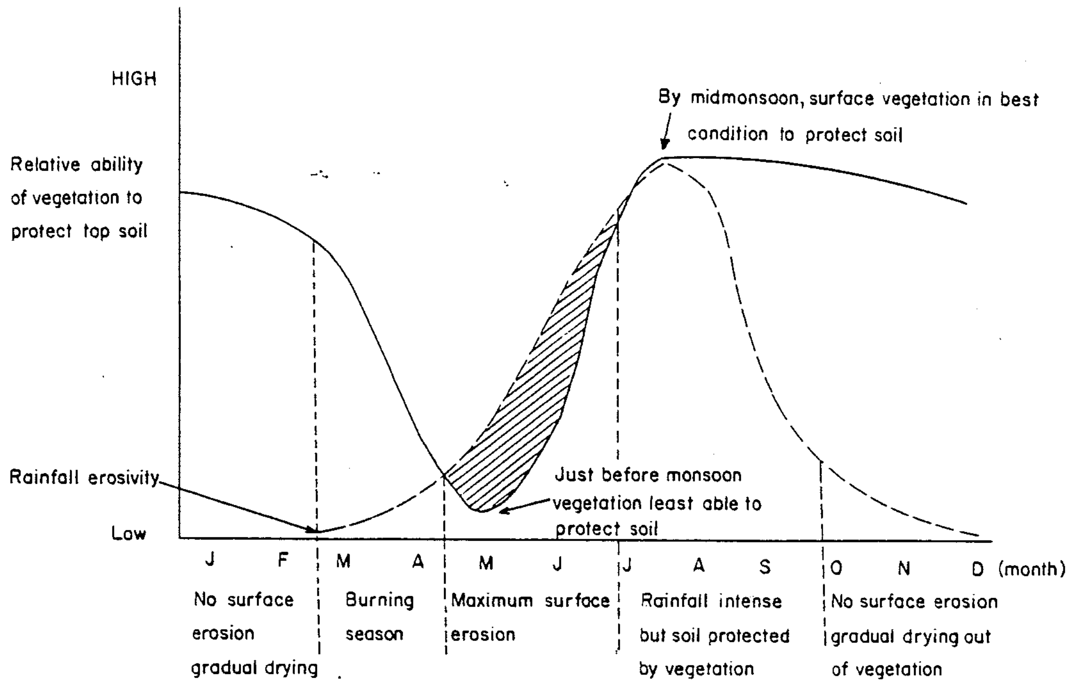


Figure 41 (a) Hypothetical course of rainfall erosive power and state of surface vegetation throughout the year in the Middle Mountains of Nepal (after Carson, 1985).

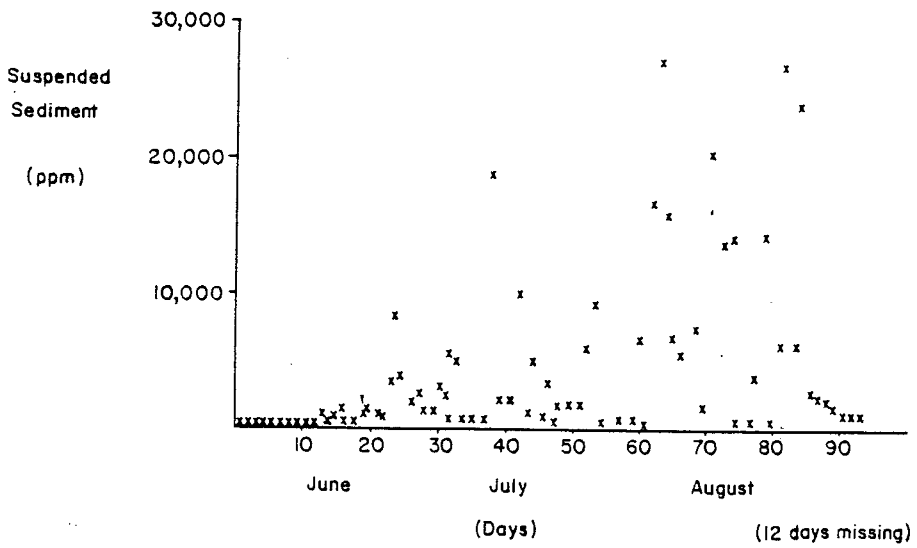


Figure 41 (b) Instantaneous suspended sediment concentrations for the Narayani river, Narayanghat, June-August, 1979 (after Carson, 1985).

sediment supply must account for the peaks observed later in the season.

Carson (1985) ascribed these variations to sudden extra inputs of large sediment amounts through landslides triggered by heavy rains, earth tremors, or both. In his own words: "One has only to observe a river system in Nepal in the late monsoon to appreciate the significance of point source sediment contribution in the Himalaya.

The major rivers are heavily sediment laden and yet the tributaries are, for the most part, sediment free. The occasional tributary, however, has extreme sediment loads that markedly discolour the main river. These tributaries are muddy because of major mass slumping into the stream, that is supplying virtually all the sediment to that river system at that time. Point source sediment contributions, caused by mass wasting, are the major contributors of sediment for many Himalayan rivers" (Carson, 1985).

As discussed earlier in Section III.5 the largest of such slides (cf. Plate 6) may temporarily block a river. When such dams burst, the result is a devastating downstream surge of water and sediment (Singh et al., 1974; Mahmood, 1987; Galay, 1987). Mahmood (1987) related how such an event, occurring in June, 1980, transported about 60 million tons of sediment in 14 hours on the Tamur river in East Nepal. This amount was equivalent to 36% of the river's annual load and five times the average load for the month of June.

Such sites often remain an active source of sediment for quite some time after the initial event, especially during higher flows, when material that was temporarily deposited as a fan at the base of the slide is again eroded (Plate 6; Carson, 1985).

Even larger amounts of debris may be carried by glacial lake outburst floods (Galay, 1987; Plate 16). One of the most spectacular of such events in historic times was the GLOF on the Seti Khola in the Pokhara valley, which occurred about 600 years ago (Yamanaka, 1982). During the event, over 5.5 km^3 of material were deposited in the valley, in which giant river terraces have since developed (Fort & Freitel, 1982; Carson, 1985).

An even larger landslide has been described by Heuberger et al. (1984). In this case, a 15 cubic kilometre chunk of a Himalayan mountain fell into the Langtang valley (North-Central Nepal) about 30,000 years ago. According to Carson (1985), many of the high river terraces in the Himalayas (cf. Plate 21) can be explained by such catastrophic events in the past, that occurred higher up in the watershed.

On a more recent time scale, Goswami (1985) reported on the effects of the August 1950 earthquake in Assam on the sediment load of the Brahmaputra in Assam Valley between 1955 and 1979. During this earthquake, apparently one of the most severe ever recorded, massive landslides occurred which temporarily blocked the Subansiri, Dibang and Dihang rivers (Figure 1).

Bursting of these dams after several days (!) not only produced devastating floods downstream, but also brought down enormous volumes of sediment, thereby raising the beds of the above rivers considerably (Poddar, 1952).

The mean annual suspended sediment load and water discharges of the Brahmaputra at Pandu (cf. Figure 2b) from 1955 to 1963 were 750.000 m^3 and 16,530 m^3/sec as against 130,000 m^3 and 14,850 m^3/sec from 1969 to 1976, respectively (Goswami, 1985).

During the former period water levels at selected discharges were observed to be rising, whereas a falling trend was noted during the latter period. Goswami (1985) concluded that the river reach upstream of Pandu was aggrading between 1955 and 1963. Similarly, the channel was degrading between 1969 and 1976.

However, one should be extremely careful with the interpretation of changes in sediment load with time. The increased sediment discharges observed in the late 1970s at Pandu, for example (Figure 42), appeared to

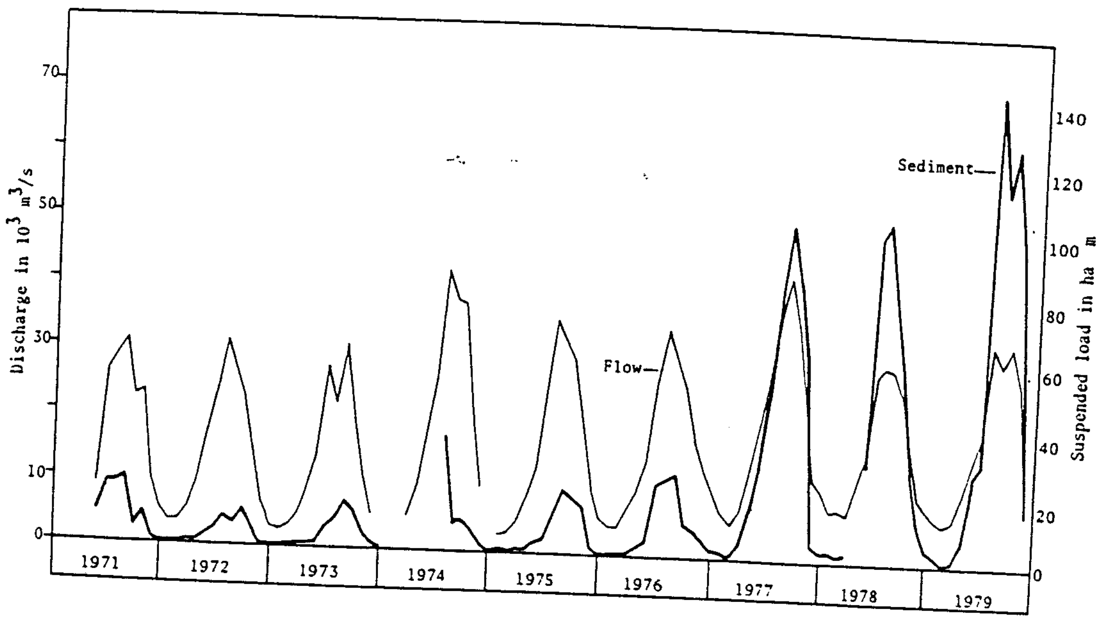


Figure 42. Mean monthly flow and sediment discharge for the Brahmaputra river at Pandu, Assam, 1971-1979 (after Goswami, 1985).

reflect temporary degradation of the reach immediately upstream of the gauging site.

The reach immediately downstream experienced significant aggradation during that time (Goswami, 1985).

As such, trends in aggradation or degradation at a particular channel cross section may not be indicative for an overall river reach. An analysis of aggradation and degradation rates for the Brahmaputra river bed over a stretch of more than 600 km and over several decades revealed the following (Goswami, 1985):

- There was a considerable gain in sediment in all reaches except one, with aggradation ranging from 0.5 to 2.4 m between 1957 and 1971 (i.e. up to 20 years after the earthquake).

- Between 1971 and 1977 an average degradation of about 20 cm was determined; i.e. only a small fraction of the material deposited earlier was again removed.

- Since streamflow amounts did not differ appreciably between the two periods, the recent removal of sediment from the river bed reflected decreased rates of sediment inputs to the system.

Thus, there appear to be phases of rapid aggradation associated with extreme events followed by periods of relatively slower removal (Goswami, 1985).

Any predictive equations of stream sediment load that fail to take into account such extreme events are bound to produce gross underestimates (Singh & Gupta, 1982). Therefore, it is no surprise to learn that rates of reservoir siltation predicted by the standard equation developed by Khosla (1953) on the basis of observations on "peninsular rivers", were severely exceeded in the Indian Himalayas (Gupta, 1983). As such, there is no need to ascribe the discrepancies between "predicted" and observed sediment loads to "deforestation" as is frequently proposed (e.g. Murty, 1985; Tejwani, 1987, and many others). This aspect will be discussed in more detail in Section IV.2.3.

IV ROLE OF VEGETATION AND LAND USE

Now that the broad regional hydrological patterns have been discussed in Chapter III, we are in a position to examine the hydrological effects of changes in land cover more closely.

As pointed out by Hamilton (1987), the term "deforestation" is often used so ambiguously that it has become rather meaningless as a descriptor of land-use change. As such it should be replaced by a more specific description of the actual activity: e.g. commercial logging of natural forest followed by replanting with fast-growing trees, or forest clearing for permanent agricultural cropping, etc.

In the following sections the effects of forest conversion to other types of land use and vice versa (i.e. reforestation) will be discussed with respect to annual water yield (IV.1.1), dry-season flows (IV.1.2) and peakflows (IV.1.3), as well as on surface erosion (IV.2.1), mass movements (IV.2.2) and stream sediment loads (IV.2.3).

IV.1 HYDROLOGY

IV.1.1 Annual water yield

A common notion about the role of forests is that the complex of litter, roots and forest soils acts as a "sponge" soaking up water during rainy spells and releasing it evenly during dry periods (Eckholm, 1976).

Although there is no doubt that forest soils generally have higher infil-tration rates and storage capacities than soils having less organic matter or that have become compacted (Patnaik & Virdi, 1962; Gilmour et al., 1987; see also the next section), tall forests also exhibit greater evapotranspiration (ET) rates than most other types of vegetation or land-use.

Evapotranspiration from forested areas consists largely of two terms, viz. transpiration (i.e. uptake of soil water by the roots, ***Et***) and rainfall interception (i.e. evaporation of water intercepted by the canopy, ***Ei***). The third term, evaporation from the forest floor (***Es***), is rather small and can often be neglected (Pathak et al., 1985)

Both ***Et*** and ***Ei*** are usually larger in the case of tall forest than for any other type of vegetation. The former because of the generally better devel-oped root systems of trees, and the latter because of the greater aero-dynamic roughness associated with tall objects (Calder, 1982).

As such, total amounts of streamflow from forested basins may be expected to be smaller than from non-forested ones, other factors being equal (Bosch & Hewlett, 1982).

What happens if the original forest cover is removed in terms of total water yield ?

When confronted with this question, one immediately faces a methodological difficulty. Simply comparing streamflow figures for adjacent catchment areas with contrasting land-use types may lead to wrong conclusions, because of possible differences in basin leakage (Section III.2).

For example, Balla (1988a) reported streamflow totals for two small (50 ha) catchments in the lower Middle Hills of West Nepal, suggesting that water yield from the forested basin might be some 130 mm/yr less than that from the adjacent agricultural basin (corrected for 28 % of trees / scrubland in the latter).

Although this value seems quite reasonable at first sight, one cannot be sure to what extent the quoted difference reflects a real vegetation effect or also a difference in catchment leakage.

Likewise, a comparison of streamflow totals for the (pine)-forested Bemunda catchment in Tehri Garhwal (Puri et al., 1982) with runoff values for the nearby agricultural Fakot area, as produced by very similar rainfall totals (Anonymous, 1984), suggests the annual water use of the pines to be lower than that of the crops by more than 340 mm/yr. This unlikely result must reflect differences in non-recorded sub-surface flow between the two areas, since the Fakot basin is much smaller (370 ha) than the Bemunda catchment (1754 ha). Similar examples from the more humid tropics have been detailed by Bruijnzeel (1989b).

Another complicating factor in the evaluation of the hydrological effects of cover transformations is the year to year variability of weather. Also, areal precipitation estimates for larger catchments, especially forested ones where raingauge densities are usually low, are frequently unreliable due to the large spatial variability of rainfall (Section II.3.1).

Zollinger (1979) examined a time-sequence of streamflow data for the Sapt Kosi in eastern Nepal for the period 1948 to 1976 and suggested that deforestation had produced a recent increase in the discharge rates of the river.

However, annual streamflow totals showed a similar trend to the corresponding rainfall figures for Chainpur East (assumed to be more or less representative of trends for the area at large) over the period 1948 - 1985 (Figure 43a). Furthermore, a comparison of the relationship between rainfall and streamflow totals for the first and second halves of the period under observation did not show a statistically significant difference (Figure 43b,c).

One might conclude, therefore, that variations in rainfall override any "deforestation" effects in the case of the Kosi. Alternatively it could be argued that the time span for which data are available is far too short to evaluate such an effect, since presumably the bulk of "deforestation" took place well before 1948 (Section II.4.3).

Similarly, Dyhr-Nielsen (1986) was unable to detect any systematic changes in streamflow patterns in eastern Thailand, despite extensive forest clearance over the last 30 to 40 years. Qian (1983) arrived at the same conclusion after analyzing streamflow and rainfall records for the island of Hainan, southern China, over the 1960s and 1970s, during which period large-scale "deforestation" occurred.

It would be interesting, therefore, to conduct a similar analysis for those Himalayan rivers for which much longer time series of streamflow measurements exist, such as the Ganges at Hardwar (gauged since 1901 (Seth & Datta, 1982), with long-term rainfall data available for 23 stations within the headwater area itself and for another 37 sites in the neighbourhood: Dhar et al., 1982b).

In fact, a rainfall-runoff analysis for the Ganges headwater basin has been conducted by the Uttar Pradesh Irrigation Research Institute for the period 1925-1956. However, no mention was made of any trends that could possibly be related to changes in land-use over time. In addition, the correlation between streamflow and rainfall was low (r = 0.44; Anonymous, 1981a).

An effective way to overcome most of the above-mentioned problems regarding the determination of the effect on streamflow of land-cover transformations is the ***"paired catchment method"*** (Hewlett & Fortson, 1983).

The technique basically involves a comparison of streamflow outputs from (at least) two basins of similar size (usually rather small), geology and vegetation. One is called the "control" (to be left unchanged throughout the observation period) and the other the "experimental" or "treatment" basin. The comparison is made during an initial calibration phase (which may take several years, depending on local rainfall variability) and during a subsequent treatment period (Figure 44). The degree to which linear regression equations (Figure 45a) or double mass curves (Figure 45b) relating streamflow totals from the two catchments (as derived during the calibration period), change after the treatment, is a measure of the effect of the latter. The total duration of such an

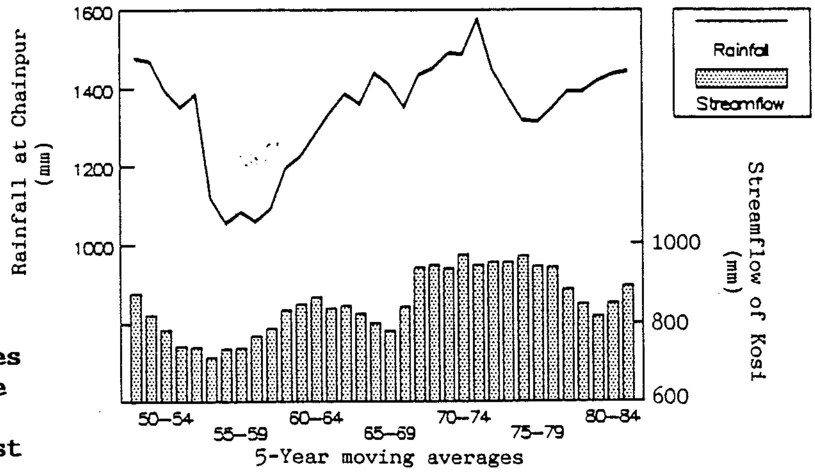
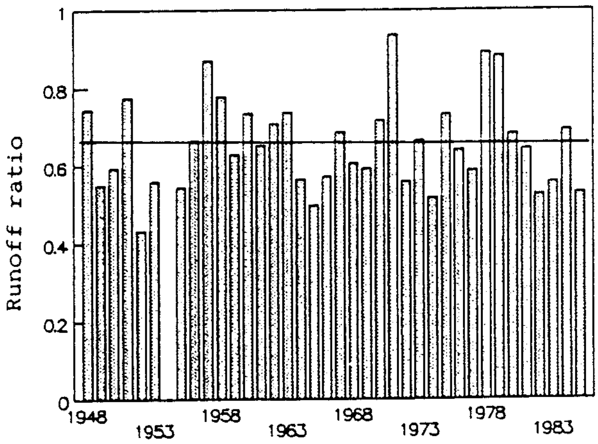
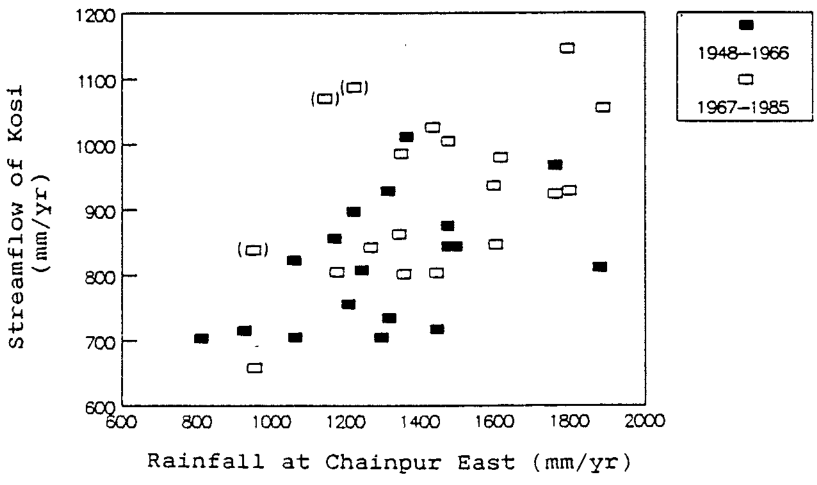


Figure 43
(a) Five-year moving averages of annual streamflow for the Kosi river at Bharakshetra and rainfall at Chainpur East (mm), 1948-1985.



(b) Annual ratio between streamflow at Bharakshetra and rainfall at Chainpur East, 1948-1985.



(c) Scatter plot of the variables from Figure 43b.

Sources of data:
streamflow 1948-1976: Zollinger, 1979;
streamflow 1977-1986: Uprety, 1988;
rainfall: Mr. Madan Basnyat, personal communication).

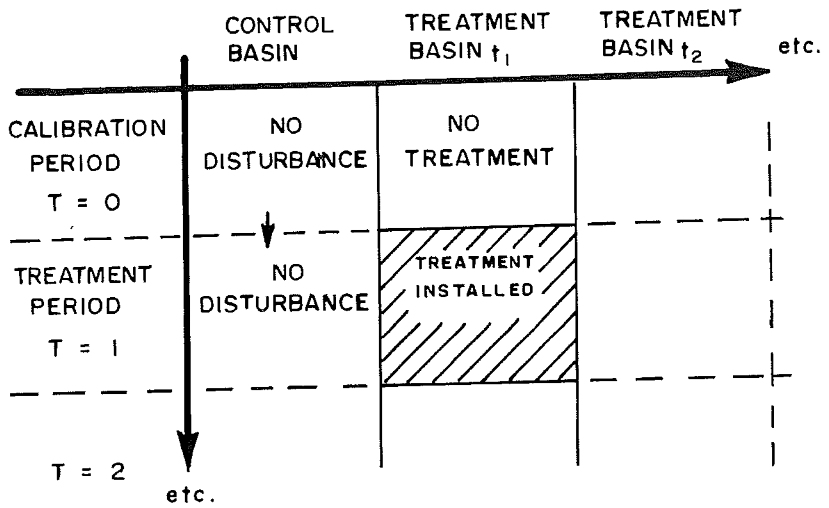


Figure 44. The paired catchment technique: general experimental design (after Hewlett & Fortson, 1983).

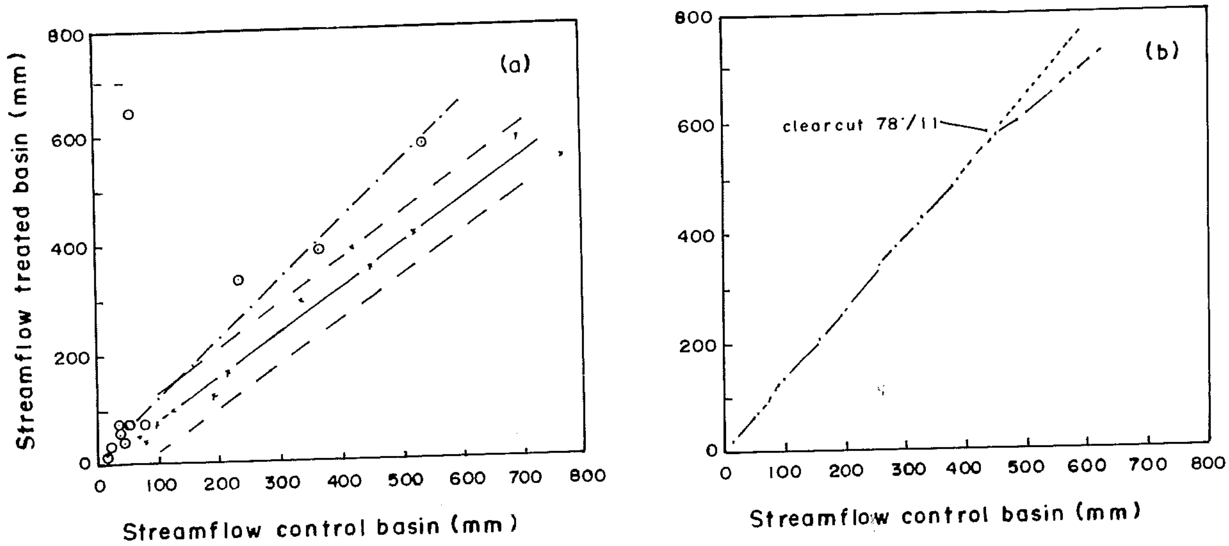


Figure 45 Evaluation of treatment effect on streamflow by statistical analysis, (a) linear regressions, (b) double mass-curves (after Hsia & Koh, 1983).

experiment may easily span a decade (calibration, clearing of forest, planting, maturation of the new vegetation).

Bosch & Hewlett (1982) reviewed the results of almost hundred paired-basin experiments throughout the world, including a few from the (sub)-tropics, whereas Bruijnzeel (1986, 1987) updated the data set for the tropics.

These authors concluded that replacement of tall forest by annual crops or other shallow-rooted vegetation invariably led to permanently increased streamflow totals, whilst reforestation of degraded lands produced a decline in total water yield.

The magnitude of the change depends not only on the type of conversion, but also on a region's climatic and geological setting as well as on rainfall patterns after the conversion (Bosch & Hewlett, 1982; Bruijnzeel, 1986). Regardless of the type of conversion, the highest increase in discharge is usually observed in the first year after forest removal, followed by a more or less regular (depending on rainfall) decline associated with the establishment of the new cover (Hibbert, 1967).

Table 7 collates the results from several paired-catchment and other studies from the Indian sub-continent. Data from a few experiments outside the region with comparable climatic conditions have been added for comparative purposes.

Although the data set pertaining to the Indian sub-continent is small, the results do confirm the general findings of Bosch & Hewlett (1982) and Bruijnzeel (1986). In other words, conversion of forest to agricultural crops leads to higher streamflow totals (Balla, 1988a), whilst replacement of grazing or cultivated land by fast-growing tree plantations produces a gradual reduction in water yield after about four years (Jaykumar et al., 1980; Samraj et al., 1988).

It should be noted, however, that the reduction in streamflow of about 20 % approximately five years after planting (Table 7), corresponds to partial conversions only (47 and 59 % of the study catchments respectively).

As such, larger reductions (35-45 %) may be expected in the case of converting entire watersheds, especially in areas of high rainfall (Jaykumar et al., 1980). Reforesting large areas in the Middle Himalaya with fast-growing exotic species will hence undoubtedly influence regional streamflow totals (see also Section IV.1.2).

The results obtained in the Selakui study (Doon Valley) have limited applicability, since the streams there are not perennial and unknown amounts of water will be leaving the catchments as sub-surface flow (see also Section III.4). They do show, however, that a well-developed understorey, once exposed to full sunlight after coppicing the main tree crop, plus the vigorous regrowth of the coppiced trees may well reduce flows considerably (Viswanatham et al., 1982). Samraj et al. (1984) also reported a 30 % reduction after coppiccing eucalypts in the Nilgiris. Since the magnitude of the reduction appears to decrease with time after coppicing (Table 7), it would be interesting to observe whether this trend will persist throughout the second rotation period of ten years. Further work on this important aspect of forest management is necessary, also in physiographic zones other than the foothills.

The preliminary work of Balla (1988a) in the Middle Hills of Nepal suggests that under these conditions a gain in streamflow of about 25 % may be expected when converting ***Shorea*** forest to well-terraced cultivated land. It cannot be emphasized enough, however, that this value was not derived through a paired catchment investigation covering several years. Again, further work of a more stringent nature is needed, if the effects of land cover transformations on water yield and timing in the Middle Himalaya are to be evaluated properly.

Table 7 Land cover transformations and changes in water yield: results from selected studies

Location and Physiographic Zone	Type of transformation	Catchment size (ha)	Elevation (m.a.s.l.)	Precipitation	Change in water yield (mm. yr) 1st year	2nd year	3rd year	nth yr	Remarks
Indian Sub-continent									
Selakui, Dun	5-yr old secondary scrub-land replaced by Eucalyptus grandis and E. camaldulensis in 30 cm deep pits (2x2 m spacing)	control 0-87 treated 1.45	ca. 520	ca. 1430 (monsoon only)	-26% over first five years				Relatively flat topography; top-soil infiltration improved because of numerous pits dug for tree planting/storey growth; flow not perennial; quoted figures correspond to summer monsoon season only and therefore mainly represent stormflow; vigorous growth of under-storey and coppiced trees
ibidem[2]	Eucalypts coppiced after 10 years	as above	as above	as above	-68%	-47%	+2%		Typical Siwalik watersheds with sandy soils; flow starts after 350 mm of cumulative rain has fallen
Rajpur Siwaliks[3]	Dense Shorea forest subjected to 20% thinning	control 6.5 treated 5.2	895	2950	no detectable change		-	-	
Gorkha, Middle Hills[4]	Single catchments with Shorea forest and agricultural (72%) plus scrub land (28%)	forest 50 cultivated 49	675 775	1670				+130 (ca 26%)	Actual recorded difference of ca 90 mm/yr corrected for 28% trees/scrubland; rainfall input determined at Ghorka, 6 km away; catchments probably watertight
Parsons's Valley, Ootacamund, southern plateau[5]	Single catchment: natural grass-land for 47% planted with Acacia melanoxylon, Eucalyptus globulus in 1960	1450	2150?	ca 2050				-290 (-22%)	Quoted value derived by inserting mean annual rainfall in pre- and post-planting regression equations (1930-1957 and 1965-1979 respectively) for single watershed
Wenlock Downs, Ootacamund, southern plateau[6]	Natural grassland plus scattered stunted evergreen montane forest planted with Eucalyptus globulus for 59% of catchment area	control 33 treated 32	2200	1535	average over 10 years: -87 mm/yr (-16%) ibidem after 4-10 years: -120 " (-21%)				Very deep soils, low rainfall intensities, no erosion or overland flow; eucalypts grown in 10-yr rotations
Elsewhere in the (sub-)tropics									
Lien-Hua-Chi, Taiwan[7]	Clearcutting of mixed ever-green hill forest; regeneration	5.9	2100	725-785	+448 (58%)	+204 (51%)			Results perhaps applicable to intermediate elevations in eastern parts of Brahmaputra basin
Kimakia, Kenya[8]	Montane rain forest/bamboo vs. plantation of Pinus patula of high stocking	36.4	2200	2440					
	I agricultural intercropping phase (3 yrs)					+127			
	II rapid growth phase until canopy closure (6 yrs)						+76		
	III continued growth; stabilised canopy (7 yrs)						+6		
Kericho, Kenya[9]	Montane rain forest/bamboo to tea plantation (54%)	702	2035	2300					
	I clearing & planting (3 yrs)					+220			
	II plantation establishment (4 yrs)							(-100)	
	III plantation maturing (6 yrs)								

[1]Mathur et al. (1976); Mathur & Sajwan (1978); [2]Viswanatham et al. (1980, 1982); [3]Subba Rao et al. (1985); [4]Balla (1988a); [5]Jaykumar et al. (1980); [6]Samraj et al. (1988); [7]Hsia & Koh (1983); [8]Blackie (1979a); [9]Blackie (1979b)

Accepting the above-mentioned values of gains and reductions in streamflow following clearing, c.q. planting of forest at face value, it would seem that the replacement of the original *Shorea* forest (as opposed to degraded scrubland) by faster growing eucalypts could also lead to a moderate reduction in water yield (Table 7). Experimental work is necessary to test this assertion, however.

The non-Indian studies quoted in Table 7 are all characterized by slightly higher rainfall totals than usually experienced in the western Himalayas. In the absence of local studies, however, the results could be used to assess the hydrological effects of such important conversions as the establishment of tea or pine plantations on former montane forest land (Blackie, 1979a,b) or temporary clearing (Hsia & Koh, 1983), e.g. in the context of shifting cultivation in the eastern parts of the basin.

There may be one exception to the general rule of increased water yield following forest clearance: the intriguing vegetation type that is often called "cloud forest" or "mossy forest" (Stadtmüller, 1987; Plate 11). In such cases, the loss of additional inputs of moisture through "cloud stripping" (Zadroga, 1981) after logging may offset the gains associated with reduced evapotranspiration (Harr, 1980). Virtually no experimental work has been published in this regard for the tropics (Hamilton & King, 1983; Bruijnzeel, 1989a). It would be wise, therefore, not to undertake any conversion of mossy forest in the Himalaya to grazing land before more is known about the possible hydrological consequences.

Although large-scale reforestation would certainly influence streamflow totals within a physiographic zone (say, the Middle Mountains), the effect could be lost at the macro-level. For example, if all cultivated and grazing land of the Nepalese Middle Mountains (ca. 15 % of the country: Carson et al., 1986) were converted to eucalypt plantation forest (assuming an eventual reduction in local flow of 35 % : Samraj et al., 1988), the effect on the total flow of the Ganges would be in the order of 3 % (section III.1), and therefore undetectable.

IV.1.2 **Dry-season flow**

Under the highly seasonal conditions prevailing over much of the region, amounts of streamflow available for irrigation or reservoir storage during the dry season become extremely important, especially during the hot pre-monsoonal months (cf. Figure 14). In this context, any further reduction in streamflow as described in the preceding section, may result in unacceptable shortages of water with disproportionately large economic losses.

Reduced dry-season flows have also oftenbeen ascribed to "deforestation" (Eckholm, 1976; Sharma, 1977; Myers, 1986). At first sight this seems to contradict the evidence presented with respect to total water yield (Bosch & Hewlett, 1982). However, the conflict can be resolved by taking into account the net effect of changes in infiltration opportunities and evapotranspiration associated with the respective land-use types (Hamilton & King, 1983; Bruijnzeel, 1989b).

Thus, if infiltration opportunities after forest removal have decreased to the extent that the amounts of water leaving an area immediately as overland flow during rainstorms exceed the gain in baseflow associated with decreased ET, then diminished dry-season flow results.

Reduced infiltration may either be caused by an increase in the area occupied by impervious surfaces (roads, villages, etc.), the use of heavy machinery during forest harvesting or subsequent agriculture, overgrazing or other improper agricultural practices (Bruijnzeel, 1986). This situation is, of course, widespread and can generally be held responsible for the commonly observed deteriora-

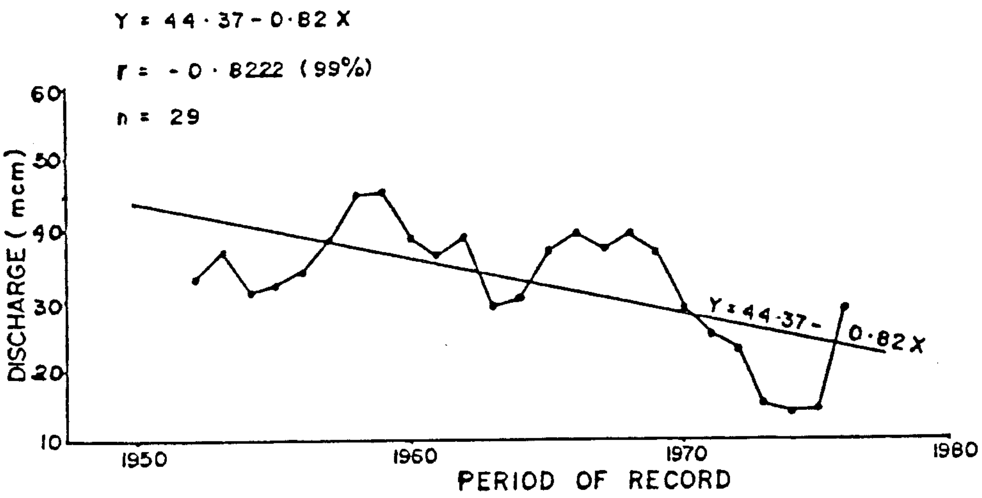


Figure 46. Decline in dry-season flow rates for the Mid-Mahaweli basin, Sri Lanka (after Madduma Bandara & Kuruppuarachchi, 1988).

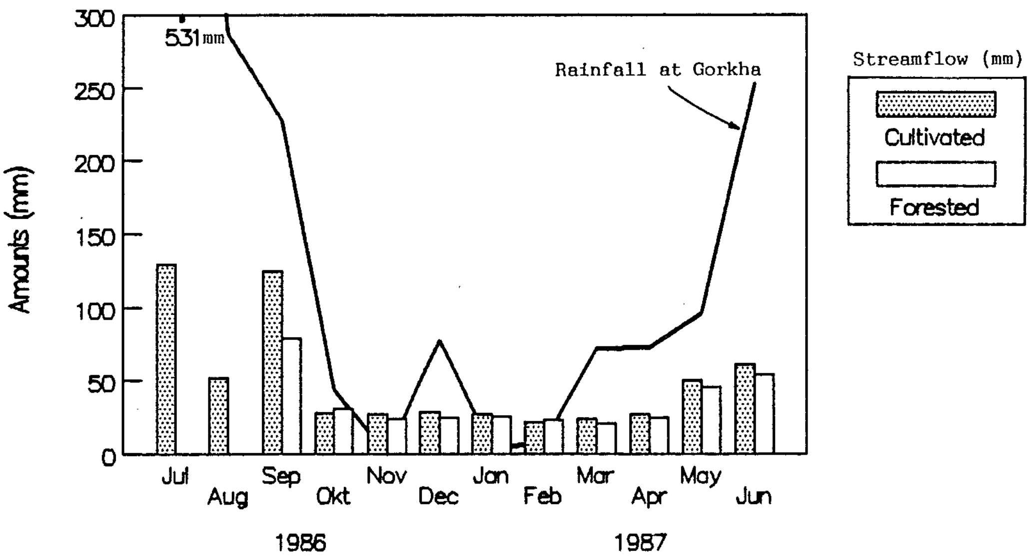


Figure 47. Monthly rainfall and streamflow for a forested and a cultivated catchment near Gorkha, Nepal (based on original data from Balla, 1988a).

tion of streamflow regimes (Bruijnzeel, 1989b). Figure 46 illustrates the point for a river basin in Sri Lanka. Here, significantly reduced low-flow rates over a period of 25 years were interpreted by Madduma Bandara & Kuruppuarachchi (1988) as being the result of the large-scale conversion of tea estates to homesteads and other crops ***without proper soil conservation measures***. Their interpretation was supported by the fact that both overall streamflow-rainfall ratios and river sediment loads had increased considerably over this period.

However, if surface infiltration characteristics can be maintained, then the effect of reduced ET after clearing will show up as increased baseflow. The effect becomes more prominent as the length of the dry season increases, reflecting differences in rooting depth between forests and cultivated crops (Eeles, 1979; Bruijnzeel, 1989b).

Figure 47 shows monthly streamflow values for the forested and cultivated catchments near Gorkha studied by Balla (1988a) (see also Table 7).

Over the (dry) period October, 1986, until May, 1987, flows recorded for the agricultural basin (including 28 % of tree- and scrubland) were generally higher than those from the forested catchment. Since the area is underlain by presumably watertight slates (Kizaki, 1987), these differences in streamflow may well reflect differences in ET between the two land-use types, though spatial variations in rainfall (e.g. in October, 1986: Figure 47) cannot be totally excluded.

It is unfortunate that Balla had to rely on rainfall measured at Gorkha town, which is not only situated 6 km from the study site, but also at a higher elevation (1100 m a.s.l. vs. 700 m). Differences in streamflow totals between the two catchments increased during wet months, though results were not very consistent (cf. September, 1986; June/July, 1987; Figure 47).

As discussed in more detail in the next section, an increased response to rainfall is generally observed following forest clearance. However, this increased stormflow is not necessarily due to reduced infiltration opportunities caused by changes in soil physical properties (which could lead to increased overland flow). It may also reflect wetter conditions in the soil associated with lower ET rates (Edwards, 1979; Eeles, 1979).

As such, the data presented in Figure 47, although perhaps not very accurate, do illustrate the point made earlier that dry-season flows from cultivated land will be higher than from forests, ***as long as infiltration capacities are maintained*** (Edwards, 1979; Bruijnzeel, 1989b).

It is unfortunate, therefore, that the Gorkha study had to be terminated in July, 1977, after only one year of observations,due to lack of funds (M.K. Balla, personal communication).

The effects on dry-season flows associated with the opposite change in land cover, i.e. the ***reforestation*** of (e.g.) grasslands, are illustrated in Figure 48 (Sharda et al., 1988). The average reduction during months of low flow was large enough (23 % at 50 % probablilty) for a partially converted catchment) for Sharda et al. (1988) to conclude that caution should be exercised when planning large-scale conversions of natural grasslands to eucalypt plantations in the Nilgiris.

Even larger reductions were observed by Samraj et al. (1984) after eucalypt coppicing. The fact that reductions in flow also occurred during the wet season (Figure 48), suggests that stormflow response was also reduced following afforestation (cf. Mathur et al., 1976). Since overland flow is a rare phenomenon in this area (regardless of land cover) due to the low rainfall intensities prevailing at this high altitude (Samraj et al., 1977; Jayakumar et al., 1978), the decreased response must reflect drier soil conditions. This was confirmed by weekly observations of soil moisture (Sharda et al., 1988).

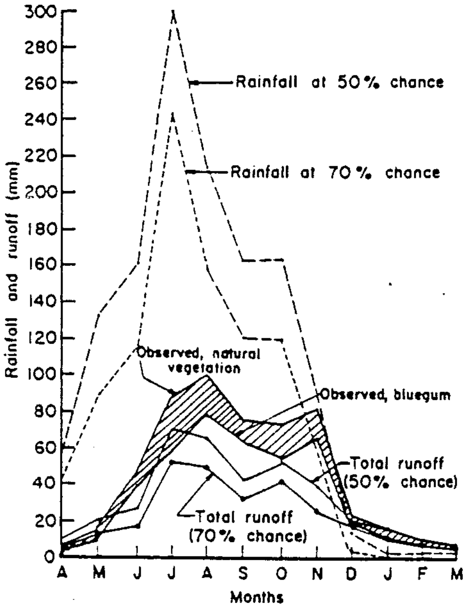


Figure 48. Expected rainfall and, total flow at 50 and 70% probability levels and monthly reduction in total flow due to afforestation with bluegum (Eucalyptus globulus) at Ootacamund (after Sharda et al., 1988).

The question could be asked, whether reforestation of severely degraded soils in the Himalayas would eventually lead to such improved infiltration conditions that the gain in amounts of infiltrated rainfall could offset losses due to increased ET.

Based on the results of Balla's study (Table 7), a value of 130 mm/yr may be taken as a first estimate of the difference in ET between (*Shorea*) forest and crops in the Nepalese Middle Hills. In the case of reforestation with chir pines, which exhibit (much) higher rainfall interception values than broad-leaved species (Dabral & Subba Rao, 1968; Pathak et al., 1985), the figure could well be twice as high, though experimental data are lacking to support this assertion.

How do the above values compare with infiltration characteristics associated with different soil- and land-use types in the Lesser and Middle Himalayas? Gilmour et al. (1987) determined the near-saturated permeabilities of top- and sub-soils under conditions ranging from heavily grazed and trampled grassland through five- and twelve-year old pine plantations to relatively undisturbed natural forest in the Middle Hill region of Nepal.

Table 8 gives a brief description of their study sites and the results of the measurements are summarized in Figures 49 and 50.

As shown in Figure 49, hydraulic conductivity values for the deepest layers of the soil profiles did not differ much between sites, since these are determined by the nature of the substrate rather than vegetative cover. The variability is seen to increase as one approaches the surface, with maximum differences occurring in the top 10 cm; average values ranged from 39 mm/hr at the overgrazed site to 524 mm/hr in the original forest (Figure 49).

Top-soil infiltration rates in the reforested sites increased with age (Figure 50). Gilmour et al. (1987) ascribed these changes to a reduction in compaction following the exclusion of grazing animals after reforestation as well as to the breakdown of litter by the soil fauna and microflora.

Before one can decide on the actual hydrological effect of such increased infiltration opportunities, one needs to take into account the prevailing rainfall intensities (cf. Figure 29). Gilmour et al. (1987) determined the average number of days on which certain rainfall intensities were recorded and compared these with

Table 8 Description of locations of infiltration tests carried out by Gilmour et al., (1987).

Site	Aspect	Slope (°)	Description
1	ENE	16	Probably deforested more than 100 years ago. Patches of surface soil exposed and surface erosion widespread. Grass cropped to soil level. Heavily grazed and trampled grassland.
2	ENE	17	Planted with *Pinus patula* five years ago. Adjacent to Site 1 and identical in condition prior to planting. Soil surface now covered with a thin layer of pine needles and short grass cover. Villagers cut grass by hand but grazing animals are excluded.
3	ESE	19	Planted with *Pinus roxburghii* 12 years ago. Similar to Site 1 prior to planting. Soil surface now covered with a layer of pine needles and patches of grass with a number of broadleaved trees developing as a minor understorey. Villagers cut grass by hand but grazing animals are excluded.
4	NE	24	Planted with *P. roxburghii* 12 years ago. Adjacent to Site 1 but the vegetation prior to planting consisted of low shrubs and herbs about 40 cm high plus a few scattered remnants of the original forest. Surface now covered with a dense layer of broadleaved shrubs and trees as an understorey beneath the pines. Leafy material cut by villagers for animal bedding and fodder. Grazing animals are excluded.
5	NE	24	Religious forest—probably a near-natural stand consisting largely of a mixture of *Schima wallichii* and *Rhododendron arboreum* with a few other minor species. Soil surface is covered with shrubs and moss but little grass. Some leafy material cut for animal bedding and fodder. Grazing animals are generally excluded.

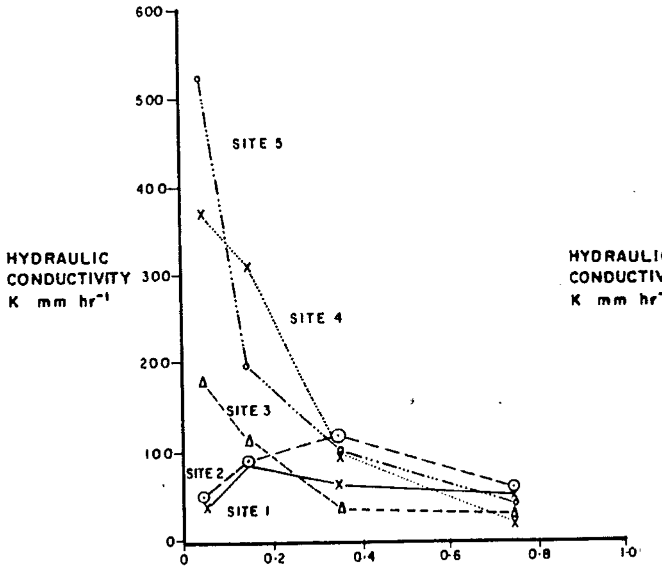


Figure 49

Field saturated hydraulic conductivity values (K) for each of the sample sites at various depths. (K values plotted against the mid-point of the sampled soil layer.)

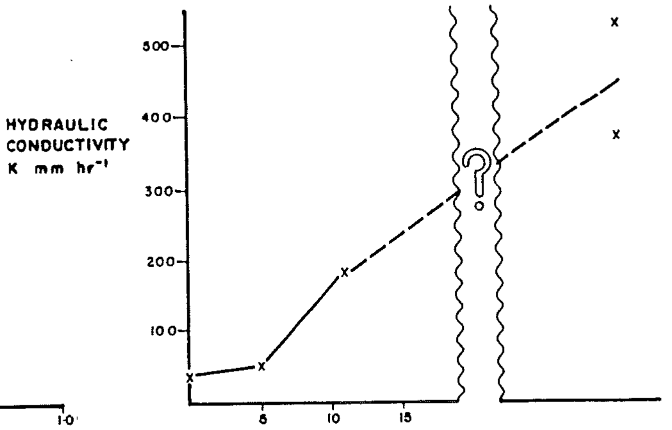


Figure 50

Likely trend of changes in field saturated hydraulic conductivity (K) of surface (0-0.1 m) soil following reforestation and protection of heavily grazed and trampled grassland.

Table 9

Annual number (and percentage) of rain-days during main monsoon season where 5-minute rainfall intensity exceeds the log-mean field saturated hydraulic conductivity for the impeding layers at each site

Depth interval (m)	Number of rain-days				
	Site 1	Site 2	Site 3	Site 4	Site 5
0 -0.1	6.7 (17%)	3.7 (9%)	tr	tr	tr
0.1-0.2	tr	tr	0.4 (1%)	tr	tr
0.2-0.5	tr	tr	6.7 (17%)	0.6 (1%)	0.6 (1%)
0.5-1.0	tr	tr	11.7 (29%)	4.4 (11%)	11.7 (29%)

the average top-soil infiltration capacities measured at the various sites. The number of occasions that rainfall intensities actually exceeded the absorption capacities of the soils was surprisingly low (Table 9). For example, even at the most degraded site, infiltration-excess overland flow would occur on average only seven times a year. At the undisturbed sites (4 and 5), permeabilities of the top 20 cm of soil were such that all rainfall could easily be accommodated (Gilmour et al., 1987).

These results naturally have important implications for the generation of peakflows, as will be discussed in the next section.

Although top-soil infiltration capacities had improved by more than 140 mm/hr, twelve years after reforesting degraded grassland (Figure 50), it is unlikely, on the basis of the relatively low rainfall intensities in the area, that the extra amounts of rain actually infiltrating the soil after reforestation will exceed the 130-250 mm/yr supposedly required for increased dry-season flow. Further experimental work is desirable.

One should be careful not to overgeneralize the findings of Gilmour et al. (1987), who themselves considered their work to be of a "preliminary nature". The interplay between rainfall and infiltration characteristics may well be different in other parts of the Himalaya. For example, Patnaik & Virdi (1962) reported much lower infiltration rates for various soils in the Siwalik-Dun zone in India (Figure 51).

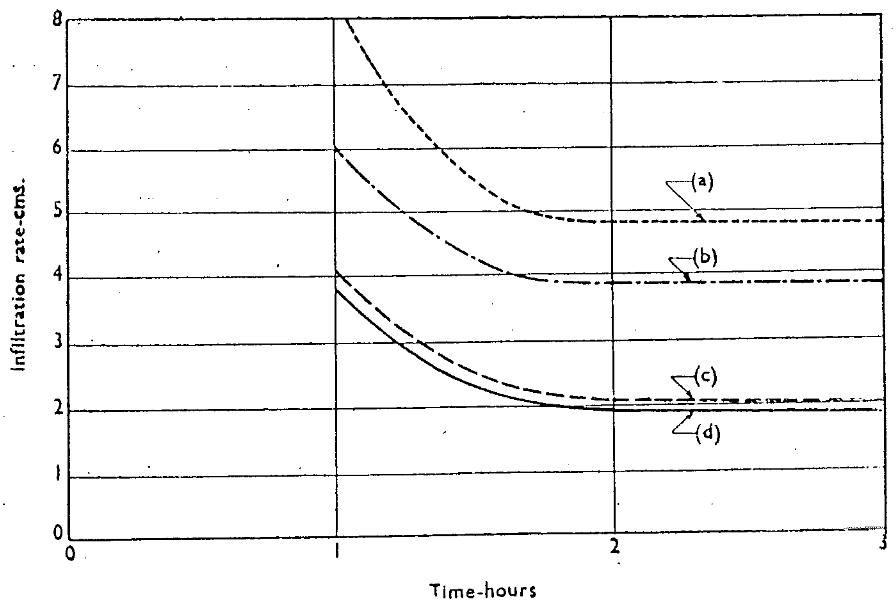


Figure 51. Hourly infiltration curves for 4 groups of sites in Doon valley: (a) cultivated Himalayan upland, (b) forested Himalayan upland, (c) forested Siwalik slope and (d) cultivated bottom land (after Patnaik & Virdi, 1962).

These investigators measured near-saturated rates for forest soils in the Doon valley of 21 mm/hr in the case of soil trampled by grazing cattle and of 58 mm/hr for "Sal forests with good leaflitter". Narayana & Sastry (1983) reported even lower values for an agricultural field in the same area.

According to various Annual Re-Reports of the Central Soil and Water Conservation Research and Training Institute, Dehradun, the number of rainstorms in the area with hourly totals of more than 30 mm range between six and eleven per year. Much higher intensities (over 100 mm/hr) occur frequently for shorter periods (Ram Babu et al., 1980).

As such, infiltration-excess overland flow is likely to occur much more frequently in the Siwaliks/ Duns than at higher elevations, especially given a situation of high grazing pressure (Pant, 1983; cf. Figure 29). Indeed, some of the most serious surface erosion in all of Nepal has been observed in the Dun zone (Nelson et al., 1980).

One can only speculate as to whether reforestation of degraded hillslopes in the Siwalik-Dun zone could restore base-flow levels. Pant (1983) maintained that prior to 1850 perennial streams rose in the hills near Chandigarh, supporting agriculture in the plains below, though he was not specific about the size of these streams.

Subba Rao et al. (1985), on the other hand, observed that streams draining densely forested small headwater catchments in the Indian Siwaliks would not start flowing until at least 350 mm of rain had fallen after the onset of the monsoon. Flow also dropped off rapidly at the end of the rainy season, possibly because of rapid infiltration into the coarse-textured riverbeds (Sharma, 1977; cf. Plates 7 and 15).

This observation was confirmed in several other small-catchment studies conducted in the area, which showed that although stormflow volumes were significantly reduced following reforestation, there was no such thing as increased baseflow (Gupta et al., 1974; 1975; Mathur & Sajwan, 1976). It must be concluded, therefore, that in this case evapotranspiration and geological effects override those associated with improved infiltration characteristics.

However, the evidence presented in this section that reforesting degraded lands will generally reduce rather than restore dry-season flows under Himalayan conditions, should not lead one to the conclusion that reforestation programmes should be discontinued. Rather, and this cannot be emphasized enough, reforestation has various on-site benefits such as reduction of surface erosion (see Section IV.2.1) and the maintenance of soil fertility, which are often more important than changes in streamflow patterns downstream (Chapter V).

IV.1.3 Peakflows

As indicated in Section III.5, there are a number of convincing reasons why regular and widespread flooding is likely to happen in the lowland areas of the Ganges-Brahmaputra river basin. It was also suggested that temporal and spatial variations in extreme rainfall constituted the chief determinant of flooding in the area (cf. Figures 32 and 33) at the meso- and macro-scales.

This section will review the evidence with respect to the influence of land management on the magnitude of peakflows and stormflow volumes at the micro-scale (up to 500 km2). In addition, we will examine the available evidence with respect to time trends.

Table 10 summarizes results obtained by several studies, mainly conducted by the Central Soil and Water Conservation Research and Training Institute, Dehra Dun, in the Indian Dun-Siwalik zone and the Nilgiris (southern plateau area).

Relatively little work seems to have been carried out in the Middle Mountain zone, which is thought to be more responsive to rainfall than any

TABLE 10. Effects of (changes in) land cover on stormflow volumes and peakflows: results from selected small-catchment studies.

Study site	Type of land use/ conversion	Effect on stormflow		Effect on peakflow	Remarks
Selakui, Dehra Dun[1,2]	-Secondary scrub vs. *Eucalyptus* sp. -Copping of the eucalypts after 10 years	-28% reduction in total flow during the first five years after reforestation -68%, 47% and no reduction		73% reduction during the first five years after reforestation	Paired catchment study in relatively flat terrain (cf. Table 7); 8 years of calibration prior to treatment; watersheds closed to grazing; streams only flow during the monsoon (June-September);
Ibidem[3,4]	-Soil conservation measures in cultivated catchment; forested basin used as control	76% reduction in total flow during the 14 years follow-wing treatment (62% in the first five years)		47% reduction for the first five years after treatment	Paired catchment study; 10 years of calibration prior to treatment in 55-ha basin; forested catchment (70 ha) heavily grazed; stormflow in forest partly from foot paths
Ibidem[4,5]	-Cultivation (no soil conservation) vs. grazed forest	Cultivated: 128 mm Forested: 166 mm		Cultivated: 3.5 m^3/sec/km^2 Forested: 6.4 m^3/sec/km^2	Long-term (1960-1983) comparison of monthly flows from single watersheds: cultivated (25 ha) and Sal forest (70 ha); quoted peakflow values correspond to a return period of two years;
Rajpur, Dehra Dun[6]	-Twenty percent thinning operation in grazed Shorea forest	Not detectable		9% increase in first year; effect disappeared after the second year	Paired catchment study in typical Siwalik environment (cf. Table 7); thinning produced an increase in crown-drip of 5%;
		1974	1975	(1974)	Multiple-paired catchment study (0.7-4.2 ha); 7 years of calibration prior to treatment in 1973; steep terrain on marls/sandstones; quoted figures computed from original data by Gupta et al.
Chandigarh[7,8]	Poorly vegetated scrub-land subjected to:				
	- annual burning	+ 26%	+ 218%	+ 225%	
	- logging + overgrazing	+178%	+ 236%	+ 52%	
	- overgrazing	+ 31%	+ 552%	+ 47%	
	- reforestation + trenching	- 75%	- 29%	- 73%	
Chandigarh[9]	Poorly vegetated grass- and scrubland subjected to reforestation with eucalypts and Acacia, contour trenches and checkdams, no grazing	60% reduction during the first 6 years after treatment		61% reduction during the first 6 years after treatment	Paired catchment study in degraded Siwalik land; 20% slope; 9 years of calibration prior to treatment in 1969
Middle Himalaya					
Naini Tal[10]	- Oak forest (47% ground-cover	0.44% of rain	-		Short-term observations of runoff from micro-headwater catchments on soil derived from limestones at an elevation of 2050 m; slopes 36-40°; stormflow values represent overland flow in a system dominated by sub-surface flow;
	- Semi-dense oak forest (41%)	0.49%			
	- Open forest (38%)	0.44%			
	- Open forest (19%)	0.45%			
	- Old landslide (31%)	0.45%			
	- Fresh landslide (18%)	0.60%			
	- Cropland (46%)	0.57%			
	- Soil deposition (12%)	0.60%			
Fakot area[11]	Cultivated part of basin (21%) bench-terraced	see Figure 52	-		Single watershed (370 ha), 36% poor woodland, 64% cultivated area plus wasteland
Lien-Hua-Chi, Central Taiwan[12]	Evergreen hardwood forest clearcut and replanted with China fir	No statistically significant effect		48% increase in median peak discharge value for three years after clearcutting	Paired catchment study in steepland conditions similar to the Middle Mountains of the Himalaya (cf. Table 7)

[1]Mathur et al., 1976; [2]Vishwanatham et al., 1982; [3]Ram Babu et al., 1974; [4]Sastry et al., 1983; [5]Ram Babu & Naranya, 1984; [6]Subba Rao et al., 1985; [7,8]Gupta et al., 1974, 1975; [9]Kausha et al., 1975; [10]Pandey et al., 1983/84; [11]Annual Reports of the Central Soil and Water Conservation Research and Training Institute, Dehadrun, 1981/1984; [12]Hsia, 1987.

other physiographic zone (Chyurlia, 1984; Section III.5).

The data collated in Table 10 clearly indicate the enormous local increases in stormflow volume and peaks that can be produced by such adverse land-use practices as burning and/or overgrazing of forest under-understorey vegetation (Gupta et al., 1974, 1975) or urbanisation/quarrying without any conservation measures (Haigh, 1982; Bandyopadhyay & Shiva, 1987) in the Himalaya.

The beneficial effects of reforesting, terracing and contour-trenching degraded soils at this scale are equally evident, as is the regenerative capacity of the ecosystem (Patnaik et al., 1974).

Although the stormflow response of a particular basin can undoubtably be modified considerably by manipulating its vegetation cover, this does not mean that peakflows from forested basins are necessarily smaller than from nearby cultivated basins.

Geological and topographical factors are also important, as is shown by a comparison of runoff values for several forested and cultivated catchments in the Doon valley (Table 10). For example, runoff from a 70-ha forested basin was consistently higher than for a nearby agricultural basin of 55 ha (168 vs. 132 mm respectively; Sastry et al., 1983).

Similarly, Pandey et al. (1983/4) found monsoonal runoff totals for very small catchments (30-240 $m2^)$ with strongly contrasting land-use types in the Middle Mountains of the Kumaon Himal to vary very little (Table 10). Since these plots were underlain by limestones, their hillslope hydrological behaviour was dominated by sub-surface flow (Figure 28).

As such, it would be unsound to directly extrapolate the results obtained in the Doon valley catchments or the plots on limestones to the Himalaya at large, with its often much steeper topography and/or different geology.

Nevertheless, dramatic reductions in monsoon runoff following terracing were also reported by Narayana (1987) for the 370-ha Bhaintan watershed in the Middle Mountain zone of Tehri Garhwal. This area, which is drained by non-perennial streams, consists of two-thirds agricultural and waste-land with the remainder under poor woodland (Tejwani, 1985).

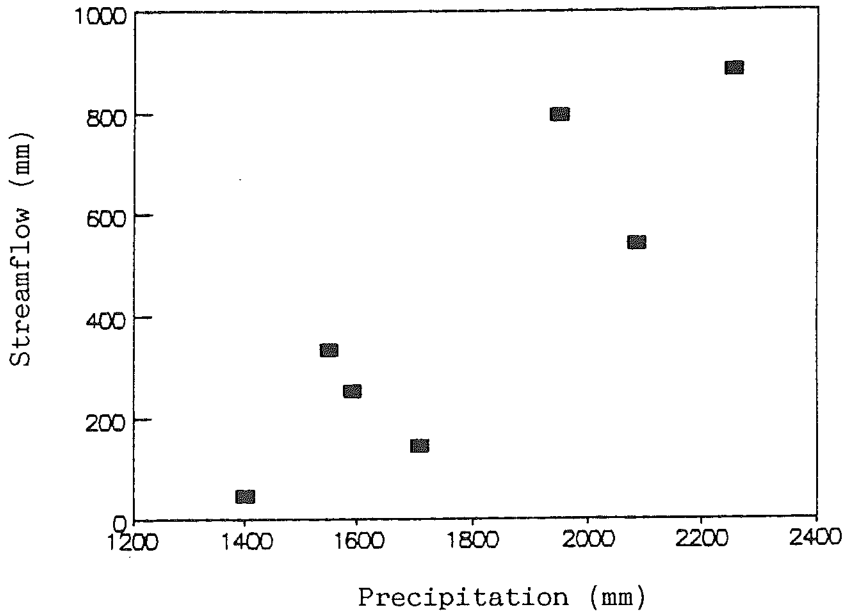
As shown in Figure 52, the lower runoff observed after terracing the fields must also be due to a reduced precipitation received in the last few years. Theoretically, the reduction over time in total streamflow for years of comparable rainfall (Figure 52) can only be explained by a concurrent decrease in overland flow from the treated fields. According to data presented in the Annual Reports of the Central Soil and Water Conservation Research and Training Institute from 1979 to 1981, overland flow from the improved terraces ranged between four and 12 % of incident rainfall, as compared to three and 34 % for unimproved terraces.

Since the corresponding runoff ratios for the catchment as a whole varied between 22 and 41 %, the flow recorded at the basin outlet must have included an unknown contribution from sub-surface stormflow (cf. Figure 31). Since terracing may be expected to increase sub-surface contributions at the cost of overland flow, it is difficult to explain the very low streamflow figures observed in some years (e.g. 1982/83; Figure 52).

A similar case of drastically reduced "runoff" following terracing 43 % of the 19.4 km^2 Coonoor watershed in the Nilgiris was presented by Jayakumar & Seshachalum (1984). Unfortunately, the results were again strongly influenced by (very) low rainfall totals following treatment.

Clearly, such uncertainties of interpretation can only be resolved by rigorous paired-catchment studies, supplemented with process-oriented hillslope hydrological work.

It is of interest to examine at what scale, land-use effects become "overshadowed" by effects associated with rainfall distribution.



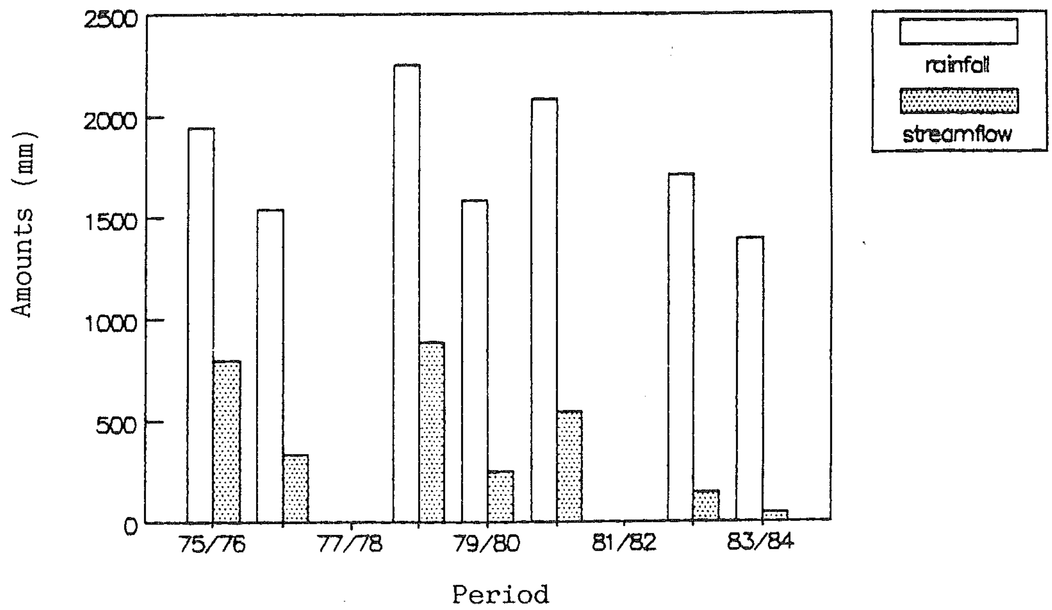


Figure 52. Annual rainfall and streamflow totals (mm) for the Bhaintan watershed, Teri Garhwal (based on data from Anonymous, 1981b/84).

The available evidence is rather limited, however. Chinnamani & Sakthivadivel (1985) described an annual time series of rainfall and streamflow data for a 43 km^2 catchment in the Nilgiris under pro- gressive urbanisation. Over the period 1949-1979 rainfall reduced by less than 10% on average. Reductions in dry-season flow were reportedly larger than that and were apparently related to decreased infiltration opportunities associated with the urbanisation process (see also Figure 46).

Although this explanation seems quite plausible at first sight, the reduction in dry-season flow may well have other causes, such as increased extraction for domestic or industrial uses, since ***total*** streamflow amounts also decreased over the period analyzed (see Section IV.1.1). Annual peakflows did not show any trend either. However, the maximum peakflow observed in 1978 far exceeded all previously recorded events (Chinnamani & Sakthivadivel, 1985). This extreme event did expose the potentially disastrous consequences of poorly planned land use under a highly seasonal rainfall regime (see also Haigh, 1982).

Sharma (1977) has described several cases of highly disturbed flow regimes following deforestation in catchments with areas of several hundred km^2 in the Nepalese Siwaliks (cf. Figure 20d).

A similar study by Madduma Bandara & Kuruppuarachchi (1988) for a river basin of about 550 km^2 in Sri Lanka also revealed a clear change in river regime following the widespread conversion of tea estates to annual cropping without soil conservation measures (Figure 46).

Garczynski (1982) drew attention to several cases with detectable downstream discharge changes (mainly in North America) where areas of more than 2000 km^2 of forest had been destroyed by hurricanes or insects.

No such trend analyses seem to be available for Himalayan basins. Although long-term time series of annual peak flows have been studied for several rivers (Anonymous, 1981a; Sakthivadivel & Raghupathy, 1978), no systematic increases in annual peakflows over time have been reported.

In fact, the data all appeared to fit such standard distributions as the Gumbel extreme value or log Pearson type III (Figure 53a). Simmilar results were obtained from analyses of annual maximum flows on the Ganges at Farakka (Figure 53b; cf. Figure 43 and Alford, 1988b) and various rivers in Assam (Jakhade et al., 1984).

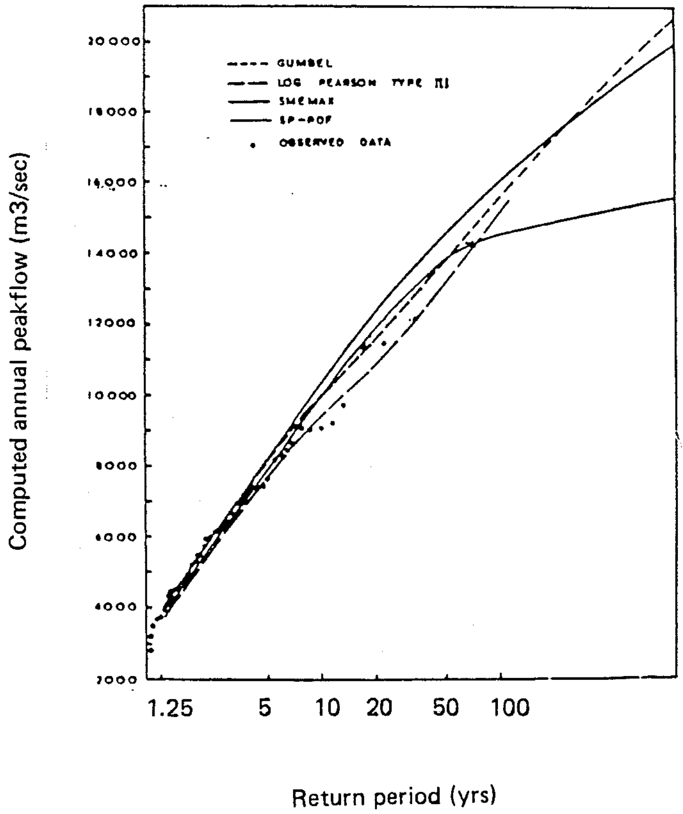
It is highly unlikely, therefore, that recent changes in land-use in the Himalayan uplands have influenced annual peakflows to any detectable degree. Again, it would seem that the timing and magnitude of extreme rainfall events are of overriding importance in this regard (Hewlett, 1982).

Although trends of increased total water yield or annual maximum flows per se have not so far been demonstrated for major Himalayan rivers, there is an undisputed increase in the area affected by floods in the lowlands since the 1950s (Rao, 1975; Bowonder, 1982).

Since it is difficult to assess the size of a flood unambiguously (stage, duration, amount of water involved?), the measure often used for policy making is the economic loss associated with a particular flood.

As pointed out by Rogers (1988), equating economic loss with severity of a flood produces a serious bias in that it gives the impression that floods are becoming more frequent and more damaging, whereas in reality increased losses mainly reflect economic growth and increased floodplain occupancy (Hamilton, 1987). As such, a flood may nowadays cause much more damage than a similar-sized one in the past (cf. Figure 18).

It is therefore understandable, however unfortunate, that many (e.g. Eckholm, 1976; Bowonder, 1982; Murty, 1985; Myers, 1986, etc.) have confused these two ways of expressing flooding severity.



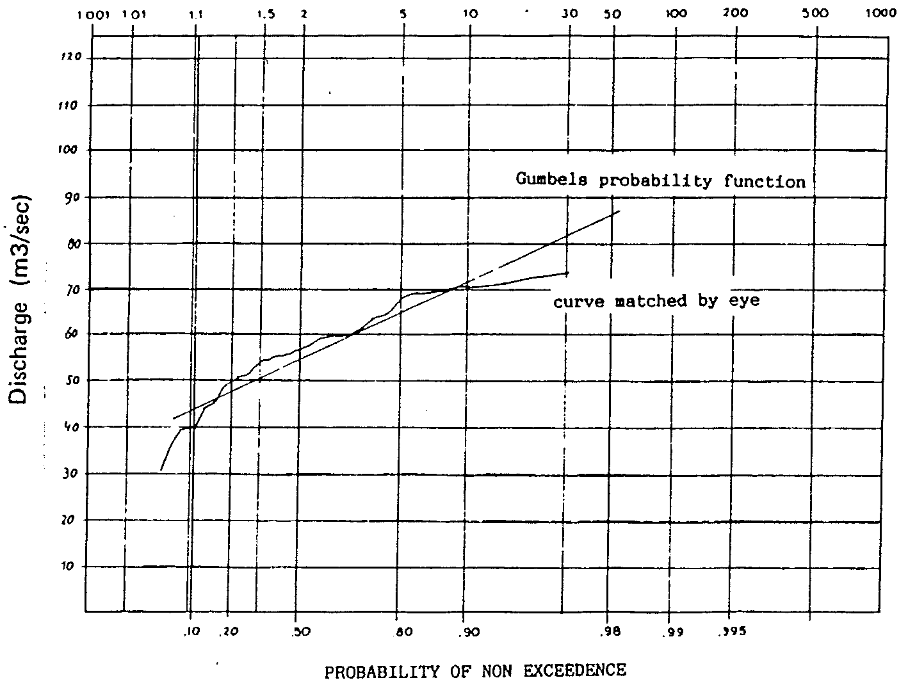


Figure 53. Annual peak discharges for different return periods for the river Ganges. (a) at Raiwala (after Sakthivadivel & Raghupathy, 1978). (b) at Farakka (based on UNESCO, 1976; Rodier & Roche, 1984).

IV.2 EROSION AND SEDIMENTATION

When dealing with the effects of changes in land use on erosion and sedimentation, it is helpful to distinguish between surface erosion (splash, sheet and rill erosion), gully erosion, and several forms of mass movements, since the ability of a vegetation cover to control the various forms of erosion is rather different.

It is equally important to make a distinction between on-site erosion (i.e. on the scale of a field or a hillslope) and off-site/downstream effects. Only part of the material eroded from a hillside may enter the drainage network, the rest may move into temporary storages in depressions, footslopes, small alluvial fans (cf. Plate 6), or be deposited in the beds of ephemeral tributary drainages or behind debris. This stored material may be transported again during large storms or become colonized by vegetation and form a stable topographic element for decades (Hamilton, 1987).

As the number of storage opportunities tends to increase with catchment size, the ratio between on-site erosion and amounts of sediment carried by a stream (the "sediment delivery ratio ratio") decreases markedly for large river basins (Walling, 1983). As we shall see later (Section IV.2.3), this scale effect has profound consequences for any off-site benefits that may be expected from reforestation programmes (Hamilton & Pearce, 1987).

Naturally, effects of erosion will be felt much more quickly on-site than further downstream. For example, soil losses from his fields may lead to such a decrease in productivity that a farmer is forced to abandon them (Shrestha, 1988), although this sediment may hardly show up in the streams of the area. Similarly, it may take decades before reduced surface erosion in upland areas is reflected in reduced sediment downstream (cf. IV.2.3).

In the following sections we will discuss the effects of changes in vegetation and land-use patterns in the Ganges-Brahmaputra River Basin on surface and gully erosion rates (IV.2.1), on mass wasting (IV.2.2) and on stream sediment loads (IV.2.3).

IV.2.1 Surface and gully erosion

Wiersum (1984) reviewed results from about 80 studies of surface erosion in (sub-)tropical forest and tree-crop systems (Table 11).

Table 11. Surface erosion in tropical forest and tree crop systems (t/ha/yr; Wiersum, 1984)

	Min	Median	Max
1. Natural forests (18/27)*	0.03	0.3	6.2
2. Shifting cultivation, fallow (6/14)	0.05	0.2	7.4
3. Plantations (14/20)	0.02	0.6	6.2
4. Multi-storied tree gardens (4/4)	0.01	0.1	0.15
5. Tree crops with cover crop/mulch (9/17)	0.10	0.8	5.6
6. Shifting cultivation, crops (7/22)	0.4	2.8	70
7. "Taungya" (2/6)	0.6	5.2	17.4
8. Tree crops, clean-weeded (10/17)	1.2	48	183
9. Plantations, litter burnt or removed (7/7)	5.9	53	105

* (a/b) a = number of locations
b = number of "treatments"

Although the data collated in Table 11 are of variable quality and reflect a variety of pedological situations, they clearly show that surface erosion is minimal in those ecosystems where the soil surface is adequately protected by well-developed litter- and herb layers (no's 1-4).

Whilst erosion rates may increase only slightly upon removal of the understorey (no. 5), it rises dramatically when the litter layer is destroyed or removed (no's 7-9). The initial effect is rather small due to the effect of residual organic matter on soil aggregate stability and infiltration capacity (no's 6 and 7), but repeated disturbances, such as burning or frequent weeding, have much more serious consequences (no's 8 and 9 in Table 11).

Incidently, this suggests that the protective value of tree stands lies not so much in the ability of the tree canopy to break the power of raindrops, but rather in developing and maintaining a litter layer (Dalal et al., 1961; Wiersum, 1985). Indeed, several recent studies have shown that the erosive power of rain dripping from forest canopies in the tropical and warm-temperate parts of the world may be substantially larger than for rainfall in the open, reflecting the larger drop size of canopy drip (Mosley, 1982b; Wiersum, 1985; Vis, 1986; Brandt, 1988).

As long as the complex of litter, herbs and understorey remains relatively undisturbed, it is able to deal with this increased striking force quite effectively, but, as already indicated, its removal may create problems (Dalal et al., 1961; Wiersum, 1984).

Results from a number of selected surface- and/or gully erosion studies conducted in and around the Ganges-Brahmaputra River Basin are presented in Table 12. The data have been grouped according to the major physiographic zones distinguished in Section II.2.

In contrast to most of the Indian studies quoted in Table 12, which often lasted for several years and may be considered to be relatively relialiable, the few data available for Nepal derive from short-term and rather site-specific studies, and do not include measurements of soil loss from cultivated land (Ramsay, 1986).

This is all the more unfortunate in view of the already indicated decrease in soil productivity in parts of the Middle Mountain zone (Shrestha, 1988). Thus far the available information on soil loss from agricultural fields in Nepal is limited to estimates based on Wischmeier's "Universal Soil Loss Equation" (e.g. Fetzer & Jung, 1979; Balla, 1988b).

Although the information collated in Table 12 concerns an array of environmental conditions, ranging from the hot and humid Meghalaya plateau with its highly erosive rainfall pattern (Figures 8 and 13) and shifting cultivation practices, to the cool and rather dry Solo Khumbu area in East Nepal where alpine grazing prevails, the data do confirm the general conclusions derived from the pan-tropical data set of Table 11.

Within each physiographic zone, surface erosion rates are generally influenced most strongly by the status of the surface, rather than by slope angle or soil type (see also Chakrabarti, 1971).

By far the highest rates of erosion seem to be associated with the second year of the cropping phase of shifting cultivation (no soil conservation practices; no's 19-22 in Table 12; cf. Hurni, 1982), and with heavily grazed areas (Plate 18), be it in the Siwaliks (no's 7 and 9), the Middle Hills (no's 13 and 14), or the High Mountain zone (no. 15).

On the other hand, erosion rates from well-kept grassland (no's 1-3, 14, 16), moderately grazed forest or scrubland (no's 6 and 14), and well-terraced (Plate 19) or mulched agricultural fields (no's 1-3, 8, 10) are low to moderate.

Gully erosion within the Ganges River Basin is especially widespread in the westernmost Siwaliks (Pant, 1983) and along the Yamuna (3900 km^2) and Chambal (ca. 5000 km^2) rivers,

Table 12. Effects of land cover on surface and/or gully erosion rates in the Ganges-Brahmaputra River Basin

Location	Type of land use or conversion	Runoff (% of rainfall)	Erosion (t/ha/yr)	Remarks
(a) **Himalaya**				
Dun valleys and adjacent slopes				
Dehradun [1,2]	- Cultivated fallow	69	185	Standard runoff plots (22.1x1.8 m); 8% slope; silt loam; maize cultivation; values are for 1978 monsoon season (1906 mm of rain); strip tillage: only along seed lines
	- Normal tillage	52	52	
	- Ibidem + mulch @ 4 t/ha	25	9	
	- Strip tillage	50	40	
	- Grass (type not specified)	2	0.4	
	- Maize, planted up and down the slope	43	22	Same plots as above; values are averages for 1982 and 1983; corresponding rainfall totals 836 and 1095 mm resp.; mulch consists of grass cuttings
	- Ibidem along centour	37	18	
	- Maize + mulch @ 2 t/ha	18	7	
	- Maize + mulch @ 4 t/ha	15	5	
Dehradun [3]	- Strawberry + weeds	28	9	Runoff plots 20x5 m; 11% slope; silt loam; values are for 1974 monsoon season (1180 mm); weeds removed by cultivation; results for clean-weeded strawberries biased due to high mortality of plants
	- Ibidem - weeds	29	26	
	- Pineapple + weeds	9	3	
	- Ibidem - weeds	11	11	
	- Pomegranate + weeds	12	3	
	- Pomegranate - weeds	31	19	
	- Perennial grass (*Cymbopagon citratus*)	20	4	
	- Cultivated fallow	25	33	
Dehradun [4,5]	- Cultivated (W3A)	9	2.1	Small catchments drained by stabilized gullies; W3A, B gauged since 1960; W3A (55 ha) bunded in 1970; post-bunding period 1970-1980; W3B (70 ha) intensively grazed; most sediment coming from trails and bank collapse; W2C (4.4 ha) gauged since 1972; checkdams erected in 1976; most sediment coming from unmetalled road; all sediment essentially bedload trapped behind weirs, suspended load unknown
	- Ibidem, bunded	6	0.1	
	- Grazed forest (W3B)	18	1.4	
	- Grazed forest (W2C)	17	4.3	
	- Ibidem, with checkdams	19	2.6	

Siwaliks				
Rajpur, Dehradun [6]	- *Shorea* coppice forest (20 years old)	42	0.2	Typical grazed Siwalik headwater catchment; 6,5 ha; sandy; 22-30° slopes; annual rainfall 2950 mm; values are averages for 1967-1970 monsoons (June-October); sediment is fraction trapped behind weir
Nurpur, H.P. [7]	- Heavily grazed forest	?	22	No further details given
Chandigarh [8]	- Terraced land, upto 50% maize cultivation	27 27	2.5 2.5	Small basin (4.6 ha); grassed waterways; values are averages for 1977-1984 monsoons (630 mm of rain)
Chatra, East Nepal [9a]	- Various, forest to grazing	-	8-37	Southerly aspect; sandstone; no further details
Surkhet, West Nepal [9b]	- Severely degraded, heavily grazed forest, on intensively gullied bedlands	-	200	Southerly aspect; sandstone; 60% slope; no further details
Middle Hills				
Fakot, U.P. [10]	- "Improved" terracing			
	1979 (1660 mm)	7	4.4	Plot of 240 m^2; terraces 2-5% lateral slope, 1-2% inward slope; plot heavily mulched in 1981
	1980 (1770 mm)	12	1.3	
	1981 (1160 mm)	4	0.2	
	- Poor terracing			
	1979 crops	23	2.8	Plot of 310 m^2; terraces 5-10% lateral slope, 5-20% outward slope; overall slope 55%; plots located in experimental watershed of Figure 52
	1980 crops	34	4.9	
	1981; natural grasses	3	2.1	
Naini Tal [11]	- Oak forest (47% cover)	0.44	0.025	Micro-headwater catchment areas of 30-240 m^2; 36-40°; elevation 2050 m; annual rainfall 2820 mm; well-drained soils derived from Krol limestones; values are averages for monsoons of 1981 and 1982 (1438 and 1561 mm)
	- Open forest (38%)	0.44	0.040	
	- Open forest (19%)	0.45	0.026	
	- Old slide (31%)	0.45	0.042	
	- Fresh slide (18%)	0.60	0.062	
	- Cropland (46%)	0.57	0.064	
	- Soil deposition (12%)	0.60	0.081	

Mussoorie [12]	- "Forested" - "Deforested"	- -	0.75 4-5	Four micro-catchments (0.8-1.0 km^2) on south-facing slope (20-30°) underlain by shales and sandstones; elevation ca. 2200 m; values are for 1978 monsoon, which was of twice-normal intensity; sediment measured trapped by road crossed by the "experimental" streams; original volumes converted to weight by application of a density factor of 1.4 g/cm^3
Banpale, Pokhara, Central Nepal [13]	- Fenced pasture - Unfenced grazing land	- -	9 35	One 10x1 m plot on south-facing stage (25°); elevation 1500 m; clay loam on phyllitic schist; four measurements between 29 June and 5 July 1978
Ibidem [14]	- Tree seedlings in fenced pasture	6-15	1	Location as above; number of plots doubled; daily sampling between 11 June and 15 October 1979; corresponding rainfall 3850 mm
	- Overgrazed land	26-47	8-12	
Tamagi, Pokhara, Central Nepal [14]	- Dense forest	<1	0.4	One 10x1 m plot on north-east facing slope (70%); elevation 1800 m; clay loam on (quartzite) schist; 11 composite measurements between 1 July and 7 October 1979
High Mountains				
Namche Bazar Dingboche, East Nepal [15]	- Sub-alpine forest - Shrub-grassland - Heavily grazed alpine	- - -	0.5 g/week 0.9 g/week 23 g/week	Thirty-five unbounded plots with 0.5 m long collection troughs ("Gerlach" type); elevations 3390 to 4416 m; sandy loams; replicated and stratified sampling design; weekly measurements between 6 April and 1 November 1984; forest on northeast-facing and scrub on south-facing slopes
(b) **Indo-Gangetic Depression**				
Agra [16]	- Cultivated fallow - *Cenchrus ciliaris* grass - Millet, cowpea, etc.	50 15 30-38	16 2 4-6	Bounded runoff plots on 2% slope; values are averages for 1981-1983 summer monsoons; corresponding mean rainfall 490 mm

Jhargram, West Bengal [17]	- Bushy *Shorea* coppice	2.5	0.31	Plots of 60x5 m laid out on 1.5% slope in *Shorea* plantation; daily measurements; plots protected from grazing since start of experiment in 1964/65; trenches 10 m apart; values are averages for 1965, 1967 and 1969; corresponding average rainfall 1463 mm; both runoff and erosion in the well-vegetated plots decreased as time progressed, whilst they increased in the burnt and bare plots
	- Ibidem, with trenches	2	0.26	
	- Tall *Shorea* coppice	2.2	0.38	
	- Ibidem, with trenches	1.5	0.22	
	- Tall coppice, burnt twice a year, no undergrowth	5	0.71	
	- Bare plot	13	4.0	
(c) Old Southern Plateau's				
Hyderabad [18]	- Broadbed-and-furrow cultivation (two crops)	14	1.6	Small catchment comparison in the black soil belt; 0.5-3.0% slopes; heavy clays; values are average seasonal (June-October) totals between 1973 and 1978
	- Traditional cultivation (rainy season fallow + weeding, one crop per yr)	25	5.6	
Shillong, Meghalaya [19]	- Shifting cultivation	6	41	No details given, but presumably representing conditions very similar to 20, 21
	- Food crops, Puerto Rican type of terracing	15	28	
	- Ibidem, but with bottom third of slope bench-terraced	12	14	
	- Ibidem, fully bench-terraced	7	3	
	- Secondary (bamboo) forest	0.5	0.5	
	- Secondary forest plus habitation	11	19	
Ibidem [20]	- Shifting cultivation first year	-	147	Unspecified number of unbounded plots sampled "periodically"; slopes 50-60%; soil lateritic; rainfall during study year 2220 mm; no information on age of vegetation in abandoned field, probably recent; measurements must be regarded as very crude and probably overestimates
	- Ibidem, 2nd year	-	170	
	- Abandoned field	-	30	
	- Bamboo forest (8-10 years)	-	8	

Ibidem [21]	- Freshly burnt fields under cropping:			One 10x1 m plot for each treatment; steep slope; (angle not specified) on lateritic soil; elevation <1000 m?; study year 1978 very dry (1420 vs. 2200 mm of normal rainfall);
	30-year cycle	21	22.5	
	10-year cycle	24	23	
	5-year cycle	26	30	
	- Fallows (regrowth)			
	5 years old	19	1.1	
	10 years old	13	0.8	
Ibidem [22]	- Cultivated fields			Four(?) 20x2 m plots for each treatment on 40° slope; elevation 1500 m; lateritic soil; values averages for two years 1977-1979; corresponding rainfall ca. 1785 mm/yr
	10-year cycle	30	50	
	5-year cycle	33	55	
	- Fallows (regrowth)			
	1 years old	25	7.5	
	5 years old	21	3.5	
	10 years old	13	2	

[1] Khybri et al., 1978; [2] Khybri, 1983; [3] Ghosh, 1974; [4] Ram Babu et al., 1980; [5] Ram Babu et al., 1981; [6] Subba Rao et al., 1973; [7] Ghosh & Subba Rao, 1979; [8] Mittal et al., 1984; [9a] Chatra Research Centre, in Laban, 1978; [9b] K.M. Sakya, personal communication, in Laban, 1978; [10] Anonymous, 1981b, 1984; [11] Pandey et al., 1983/84; [12] Haigh, 1982; [13] Mulder, 1978; [14] Impat, 1981; [15] Byers, 1987; [16] Bhushan & Prakash, 1983; [17] Ray, 1971; [18] Kampen et al., 1981; [19] Singh & Singh (1981) in Das & Maharjan, 1988; [20] Singh & Singh, 1980; [21] Toky & Ramakrishnan, 1981; [22] Mishra & Ramakrishnan, 1983a

Plate 18 Prolonged overgrazing after forest removal produced this seriously eroding landscape on vulnerable red soils in the Kabhu Palanchok district, Central Nepal.

Plate 19 Excellent terracing of a hillslope in the Middle Mountain zone of the Kumaon Himalaya enables sustainable rice cultivation (photograph by J. Rupke).

most of it caused by mismanagement of these relatively dry and therefore ecologically sensitive areas (Haigh, 1984b).

Similarly, in the Himalaya, the clearance of vegetation and subsequent overgrazing in places where highly erodible soils (such as those found on sandstones, quartzites, deeply weathered gneisses or lacustrine deposits) occur, have lead to intense gullying (Nelson et al., 1980; Brunsden et al., 1981; see also Plate 18).

The influence of vegetation in case of actively eroding gullies is rather limited and rehabilitation schemes will often need to be supplemented with mechanical measures, such as check dams, retaining walls as well as protected waterways diverting the water from the eroding headwall (Narayana & Sastry, 1985).

Although the Central Soil and Water Conservation Research and Training Institute at Dehradun has achieved a fair deal of success in developing techniques to reclaim gullied lands in the western parts of the Ganges basin (Das, 1977; Haigh, 1984b), gully rehabilitation remains a complex and often costly affair (Hudson, 1971).

Whilst the data presented in Table 12 do indicate the very real possibility of minimizing on-site, and to a lesser extent also gully, erosion rates by proper conservation practices or reforestation, it should be realized that the amounts of sediment carried by Himalayan rivers (Tables 5 and 6; Section III.6) are much larger than most of the field erosion rates quoted in Table 12. This points again to the importance of riparian mass wasting as a source of sediment in these young mountains (Figure 41b; Meijerink, 1974; Carson, 1985).

Therefore, any discussion of the possible benefits that large-scale upland rehabilitation programmes may have for sedimentation in the lowlands, is bound to raise false expectations as long as it fails to take this aspect into account (Narayana, 1987; Section IV.2.3).

IV.2.2 Mass movements

The high incidence of landsliding in the Himalaya is immediately apparent on geomorphological maps of the area (Meijerink, 1974; Brunsden et al., 1981; Lakhera, 1982; Kienholz et al., 1983; Fort et al., 1984, etc.). Indeed, the importance of mass movement processes as contributors to the overall sediment loads of streams in the Himalaya is more or less generally accepted in the (scientific) literature on the subject (Laban, 1979; Narayan et al., 1983; Carson, 1985; Ramsay, 1986; Fleming, 1988; Euphrat, 1987).

However, in contrast to this wealth of information on spatial frequencies of mass movements, there is a paucity of data on temporal frequencies (Prasad, 1975) or the actual quantities of material moved (Starkel, 1972).

Ramsay (1986, 1987b) reviewed a number of studies on mass wasting in and around Nepal. Most authors quoted by him ascribed the high level of mass wasting prevailing in the Himalaya mainly to a combination of geological and climatic factors, and to a lesser extent to land-use factors.

Steep dip-slopes, unstable nature of rocks due to their structural disposition (e.g. degree of fracturing), depth and degree of weathering, high seismicity, and oversteepening of slopes through undercutting by rivers, ranked among the most important geological factors (Bansode & Pradhan, 1975; Suneja, 1979; Brunsden et al., 1981; Caine & Mool, 1982; Peters & Mool, 1983; Wagner, 1983; Narayan et al., 1983).

Figure 54 compares frequencies of various types of landslides for different lithologies in East Nepal (Brunsden et al., 1981). Debris slides (relatively shallow (1-3 m) and elongate (100-1000 m) masses of fine earth, frequently triggered by heavy storms) and rock slides occurred on all rock types, but were much less frequent on gneiss and quartzites than on schists, shales and phyllites.

The low frequency of mass movements on the gneisses was explained by the fact that neither the structure of these rocks nor their deep and irregular weathering did permit the development of planar slide surfaces and unstable weathering mantles. As indicated above, however, these gneisses were quite prone to gully erosion (Brunsden et al., 1981). In addition, mass movements in this area were highly concentrated in low level undercut situations, such as ravines and outer river bends (Plate 20).

Of particular interest is the study by Prasad (1975), who discussed ten years of observations of seismic activity, rainfall and landslide occurrence in part of the Kosi basin in eastern Nepal. In general, landslides were most frequent during times of both rainfall and earthquake activity (mainly in July and August). Since slides also occurred during times of low seismicity, Prasad (1975) concluded that intense precipitation and the associated saturation of soils were apparently more important than seismic shocks.

The latter contention is supported by many other observations in the region. For example, Carson (1985) related how in one area in the Middle Hills of Nepal, villagers indicated that the landslides still visible in the early 1980's had all occurred during two events of heavy rain, one in 1934 (!) and the other in 1971 (cf. Manandhar & Khanal, 1988). Similarly, Starkel (1972) found that during an extreme rainfall event of more than 700 mm in three days near Darjeeling, many new landslides were initiated and old ones reactivated. He estimated the associated erosion rate at about ten times the annual average.

Of particular importance seems to be the occurrence of preferential soil water flow paths in zones of structural discontinuity, such as faults (Brunsden et al., 1981; Ramsay, 1985; Nainwal et al., 1985-86). Where also deeply weathered and fractured, such zones often give rise to what Brunsden et al. (1981) have termed "mass wasting catchments": steep, rapidly eroding channels with active, expanding heads supplying material to the channels by a variety of mass movements.

Ramsay (1985) found such mass wasting catchments to be responsible for about 90 % of all the sediment produced by mass wasting processes in the Phewa Tal watershed in the Middle Hill zone of West-Central Nepal (see also Tang et al. (1981) and Du & Zhang (1981) for a classification of debris flow types in the Trans-Himalayan zone and information on their spatial distribution).

With such strong geological and climatic controls over mass movement processes, it is somewhat difficult to evaluate the influence of certain disturbances on land-slide frequency and magnitude in the Himalaya with any degree of certainty. Also, remaining areas of forest in the Middle Mountain zone tend to be on slopes too steep for terraced cultivation. This immediately introduces the methodological difficulty of finding comparable control sites (the forested slopes being steeper and therefore more susceptible to gravity).

For example, it was hoped that the detailed geomorphological mapping of the Kakani area north of Kathmandu would shed some light on the role of forest in preventing mass movements (Kienholz et al., 1983, 1984). Unfortunately, however, the forested areas were found on a different rock formation. As such, the lack of slides on the forested hillslopes could not be ascribed with certainty to the presence of a better vegetation cover.

Laban (1979) carried out a reconnaissance survey of slope failure intensities all over Nepal, counting the number of slides larger than 50 m^2 per linear km as seen from one side of a light aircraft flying at a constant speed of 100 miles per hour.

By necessity, the methodology was simplistic: landslides occurring on forested land (i.e with more than 50 % crown cover) were assumed to be "natural", whilst those seen on land

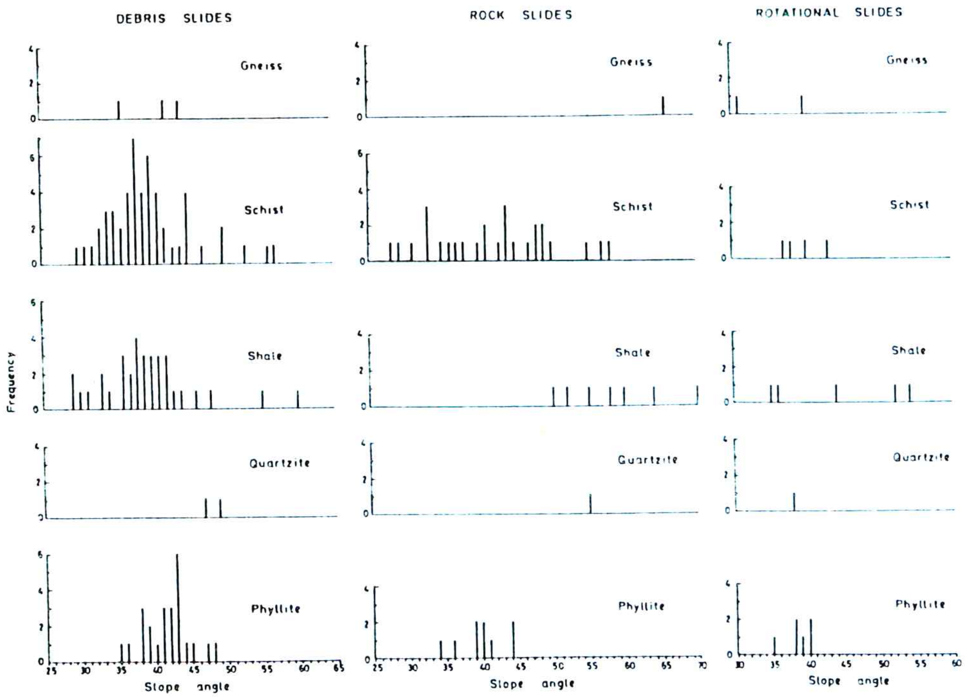


Figure 54. Frequency of mass movements as a function of slope angle and rock type in the Dhankuta and Leoti Khola areas, East Nepal (After Brunsden et al., 1981).

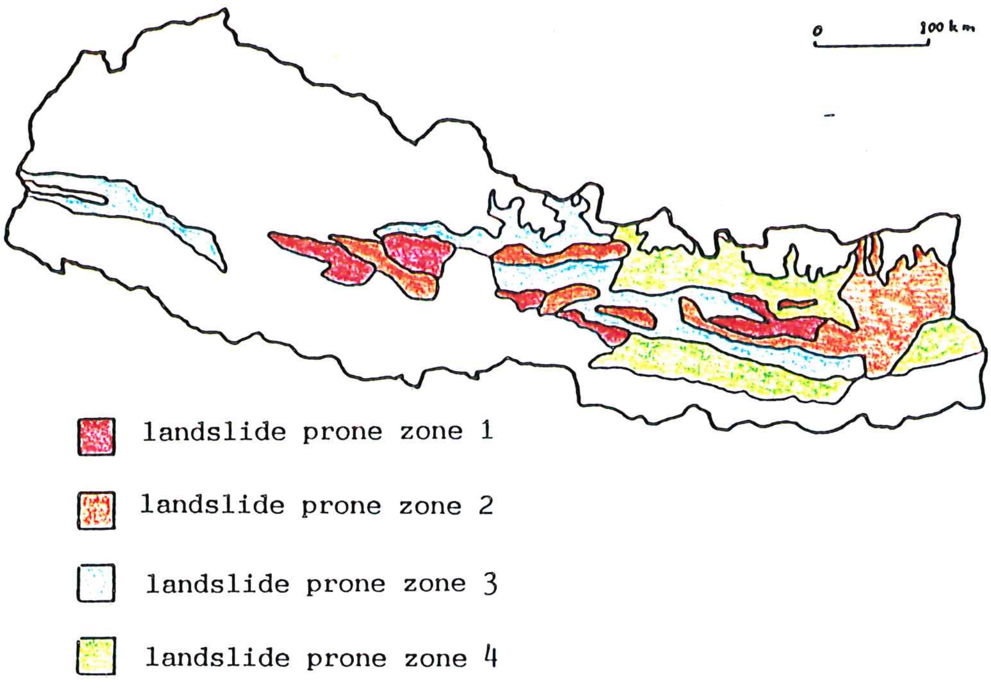


Figure 55. Preliminary distribution of landslide zones in Nepal (modified from Paudiyal & Mathur, 1985).

Plate 20

Undercutting of steep slopes in river bends constitutes a major sediment producing mechanism in the Himalaya. Note the fully forested condition of the slope.

with less than 50 % crown cover were

by implication "man-caused". Failures associated with trails and roads ("man-caused") as well as undercutting by streams ("natural") were distinguished separately.

The results, expressed per ecological zone (Nelson et al., 1980; cf. Figure 4) and combined with an index of rock erodibility, were used later as a (very rough) indicator for regional patterns of landslide hazards in Nepal by Paudyal & Mathur, 1985) (Figure 55).

Laban (1979) himself did not consider his observations "sufficiently adequate to give insight into the different roles of man and nature in precipitating landslides", although "the acivities of man increased landslide density in many cases". All in all, Laban (1979) suggested that roughly 75 % of all slope failures in Nepal could be considered as "natural". No attempt was made to convert landslide frequencies into volumes of sediment so produced.

Euphrat (1987) conducted a survey of landslide volumes, distribution and approximate age in a 55-ha catchment (47 % cultivated; 43 % scrub and grassland) near Dhulikhel in the Middle Hills of Central Nepal. He concluded that about 80 % of the 474 slides that were recorded, could be classified as "occurring naturally". More than half of these "natural" slides were encountered in the riparian zone (cf. Plates 3 & 20).

Ramsay (1986) distinguished two categories of disturbances: changes in land use (principally the removal of forest cover, followed by grazing or cultivation, possibly with terracing and irrigation), and construction activities (mainly roads, irrigation canals and housing).

As for the influence of (tall) vegetation on slope stability, the net effect is generally considered positive, the major factor being mechanical reinforcement of the soil by the tree roots (Ziemer, 1981; O'Loughlin, 1984).

Although the removal of tall vegetation may lead to wetter conditions in the soil due to reduced evapotranspiration (cf. Section IV.2), which would tend to increase slide hazard, this is not thought to be very important in the case of the Himalaya. After all, most of the failures occur during the second half of the rainy season (Prasad, 1975; Carson, 1985; see also Figure 41b), when slopes have become thoroughly wetted by antecedent rains anyway. Under such conditions, the extra cohesion imparted by tree roots may be critical to slope stability.

It is important, however, to make the distinction between deep-seated and shallow (less than, say, 3 m) slides, as the former do not seem to be influenced appreciably by the presence or absence of a well-developed root network (Starkel, 1972; Carson, 1985). Similarly, Brunsden et al. (1981) reported how mass wasting in phyllitic terrain in eastern Nepal during a few heavy storms, that occurred in late July 1974, was much more intensive on steep forested slopes than in the more gently sloping cultivated areas. Failures were generally restricted to deep ravine headwater areas and along the lower valley sides and banks where undercutting occurred.

Starkel (1972) his contention, that the role of vegetation in preventing *shallow* slope failures (often triggered during heavy rain) is "most important", was demonstrated rather dramatically by the study of Manandhar & Khanal (1988) in the Lele catchment, an area underlain by limestones and phyllites, some 20 km south of Kathmandu. Examination of aerial photographs taken in 1972 and 1986 showed an increase in the number of landslide scars from 93 to 743. Most of these failures occurred on slopes steeper than 33 degrees and had been triggered during a single cloudburst in September 1981 (see photograph III in Carson, 1985).

Interestingly, local informants told the investigators that such heavy storms occurred about once every ten to twelve years (Manandhar & Khanal, 1988), thus confirming the suggestion made by Brunsden et al. (1981) for the return period for formative events for slope and valley landforms in (East) Nepal. Only a few landslides occurred in the thickly vegetated headwater area of the catchment, the majority being found on sparsely vegetated slopes and near limestone quarries (Manandhar & Khanal, 1988; see also Haigh, 1982,1984a).

Although numerous, these small and shallow failures found in mid or upper slope positions usually heal rather quickly (Ramsay, 1985; Euphrat, 1987). In addition, they are only modest contributors of sediment to the streams as they become rarely incorporated in the drainage network (Ramsay, 1987a; cf.section IV.2.3).

Terracing of hillslopes after vegetation removal is generally not considered a direct cause of mass wasting (Ramsay, 1986), although Marston (1988) noted that poor control of terrace drainage in the Langtang-Jugal Himal area of Nepal was important in this respect. As pointed out by Carson (1985), the length and intensity of human occupation in the Middle Mountains is such, that areas liable to slide due to addition of irrigation water would probably have done so a long time ago (cf. Plate 19).

Rather, existing irrigated terraces are stable and small slumps and collapsed terrace risers (Euphrat, 1987) quickly repaired. As described by Johnson et al. (1982), farmers are well aware of increased slide hazards associated with the accumulation of water on terraces.

This perception has sometimes lead them to shift from irrigated to rain-fed cropping. A consequence of this practice, however, is an in- crease in surface erosion (Table 12), as rain-fed terraces are outward-sloping. This is done deliberately in order to dispose of excess rainfall during the monsoon, which could damage the crops by waterlogging (Johnson et al., 1982).

According to Ramsay (1986), irrigation canals in upland areas are frequently associated with slope failures due to both the removal of toe support from slopes and to saturation of the weathering mantle by seepage and overflow.

This brings us to the effects of *construction activities* on mass movements and sediment production.

The construction of large dams with the subsequent inundation of a valley will have on-site and off-site consequences. Around the reservoir itself, the increased pore pressure associated with the saturation of the slopes may well trigger slides when the water level in the artificial lake is lowered for some reason (Carson, 1985; Galay, 1987). Below the dam, a new cycle of riparian mass wasting may be initiated as the river will tend to regain its lost (i.e. trapped by the dam) sediment load by increased incision (Rudra, 1979; Mahmood, 1987; Galay, 1987).

However, by far the most important construction impact on the stability of slopes in the Himalaya, is the ***building of roads*** (Misra & Agarwal, 1982; Haigh, 1984a).

Laban (1979) estimated that ca. 5 % of all landslides in Nepal were due to trails and roads, whilst Euphrat (1987) found a comparable figure of 7 % of total landslide volume to be associated with trails in his study area in the Middle Hills east of Kathmandu. Although perhaps (still) minor in terms of total sediment contribution, this implies an enormous amount of material being moved from a limited area, presenting equally enormous problems of road maintenance (Haigh, 1984a; Carson, 1985).

Similarly, a survey of the state of the hill-roads of the Tehri-Garhwal and Dehra Dun districts of northern India, conducted at the close of the monsoon of 1975, revealed on average ten slides of 500 m^3 per km of road-bed (Bansal & Mathur, 1976).

Although it can be shown, that proper road engineering can solve many of the problems (see Schaffner (1987) his account of the building of a "green" road in the Middle Mountain zone of Nepal), it should be realized that associated costs are extremely high. For example, Carson (1985) quotes the construction costs of the Dharan-Dhankuta road in East Nepal, according to him one of the best engineered roads in Nepal, as amounting to more than 1 million US dollars per km. He went on to say that "in spite of the care made in the assessment of the alignment and during construction, this road has been closed during part of the last two monsoons by serious landslide activity. In the 1984 monsoon over 5 million dollars of damage was done by slope failure".

Clearly, the Himalayan environment is "extremely hostile to road building" (Ramsay, 1986), although the reverse is probably even more true. Therefore, also in view of the present stage of economic development of the Himalaya, the construction of single lane roads with pullouts, suited to farm tractors with trailers (rather than heavy trucks), as suggested by Carson (1985), as feeder roads in the hills holds some promise.

IV.2.3 River sediment loads

As shown in the preceding sections the influence of the presence or absence of a good vegetation cover c.q. land management on surface erosion and shallow landslides is quite pronounced at the scale of a farmer's field or very small catchments (Table 12). In this final section on land-use influences we will investigate to what extent changes in on-site erosion induced by land management show up in the sediment loads of streams draining larger catchments in the Himalaya.

Himalayan streams carry high to very high amounts of sediment, especially those in the Middle Mountains (Tables 4-6). The question may be raised to what extent this high rate of sediment production in the Middle Himalaya is man-caused. After all, the forests in this part of the Himalaya have been under rather great pressure for quite some time (Section II.4.3).

The work of Meijerink (1974) in the Aglar catchment, a typical Mid-Himalayan catchment of 320 km^2 in the western Garhwal Himalaya, is particularly relevant in this respect. On the basis of a morphometric analysis of river terrace levels, using aerial photography and field checks, Meijerink calculated that the Aglar river had removed about 1990 m^3 of sediment from a riparian zone of ca. 34 km^2 since the formation of these terraces, some 10-15,000 years ago (cf. Plates 21 and 1).

This would correspond to a long-term natural erosion rate of 40 to 60 m^3/ha/yr, i.e. similar to the values quoted for this zone in Tables 4, 5 and 6. Meijerink (1974) considered most of this sediment to have been derived from slopes that became over-steepened by undercutting by the incising river (cf. Plates 3 & 20).

It could be argued that climatic changes since the Pleistocene have been such, that erosion rates must have varied considerably. Also, the Aglar area is rather dry as it is situated in the rainshadow of the "Mahabharat" (cf. Figure 10). On the other hand, the process of riparian mass wasting is far less sensitive to climatic conditions than, for example, surface erosion.

Similarly, Brunsden et al. (1981) reported stream incision in the Middle Himalaya of eastern Nepal to have been greater than ridge crest lowering during the Pleistocene, resulting in valley-side profiles of generally convex shapes. According to the same authors, the rapidly incising river network is transmitting the effects of tectonic and iso-static uplift to the hillslopes through undercutting and landsliding, thereby moving large amounts of debris directly into the streams (cf. Plates 3, 6 and 20).

Currently accepted rates of tectonic uplift are in the order of 1 mm/ yr (Zeitler et al., 1982; Iwata et al., 1984), although rates as high as 9 mm/yr have been reported for the westernmost part of the Himalaya (Zeitler et al., 1982) and even higher values for the Nanda Devi area (J. Rupke, personal communication).

On the basis of the above evidence we may conclude that river incision is generally keeping pace with uplift and that natural stream sediment loads in the Himalaya must always have been high to very high. The very thick alluvial deposits of the Bhabar zone (Figure 6) bear testimony to this as well (Weidmer, 1981).

Interestingly, and quite unlike most other big river systems in the world, the specific sediment load (i.e. per km^2) of the Ganges increases with basin area (Figure 56). Normally, only a few sub-basins within a given drainage basin exhibit (very) high rates of sediment production (Holeman, 1968; Milliman & Meade, 1983). Not so in the Himalaya, however, where most major streams carry large amounts of sediment (Figure 37), and where the main stream flows all along the mountain range, thus receiving a continuously high input of sediment along its course (Figure 1; Abbas & Subramanian, 1984).

It will be clear from the above, that opportunities to influence such huge transport rates of sediment will be limited. In addition, sediment delivery ratio's (SDR) tend to decrease with catchment size, thereby diminishing off-site effects of changes in on-site erosion (Mou, 1986; Bruijnzeel, 1986; Hamilton, 1987), as will be demonstrated by the following examples from the Indian and Nepalese Himalayas.

Plate 21 Repeated phases of uplift and river incision produced this fine series of river terraces in alluvial deposits in the Middle Mountain zone of the Kumaon Himalaya (photograph by J. Rupke).

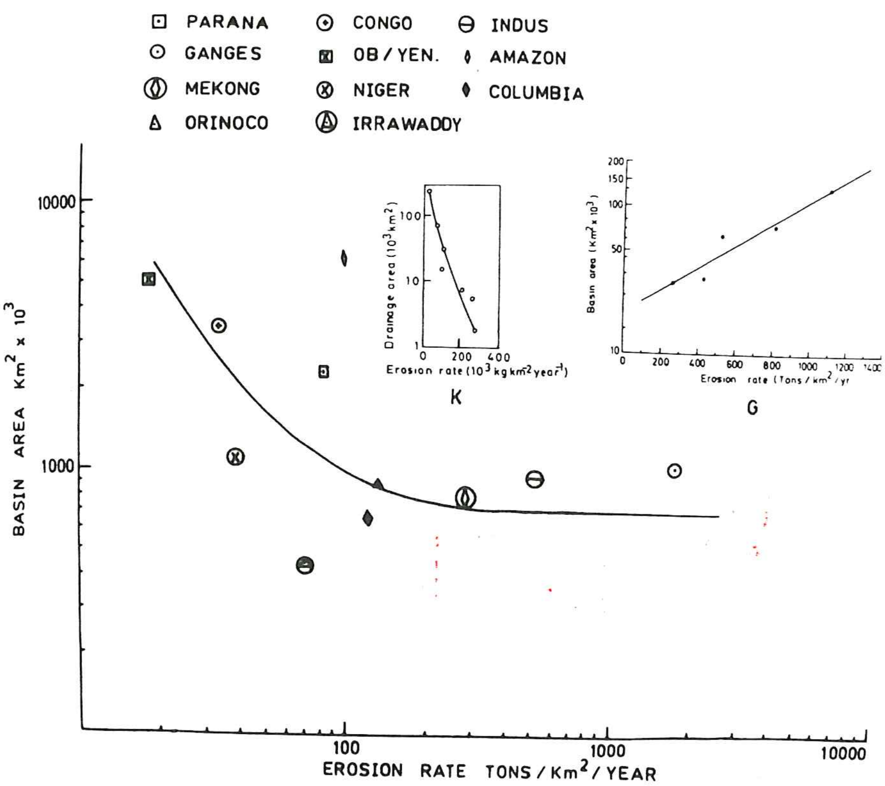


Figure 56. Sediment yield vs. basin area for major tropical river basins. Inset (G): Ganges; (K): Krishna, India (after Abbas & Subramanian, 1984).

Sukhna Lake, Chandigarh

The 229-ha Sukhna lake near Chandigarh is fed by two flashy streams of about equal size, the Kansal and the Sukhetri, which together drain an area of about 44 km^2, 80 % of which is located in the Upper Siwalik formation. This part of the basin consists of forest on steep slopes, with sandy to gravelly soils that are highly susceptible to erosion as well as rather unstable. The forest land is subjected to intense grazing and suffers from sheet erosion (cf. Table 12) and (riparian) sliding. The remaining 20 % of the basin is under cultivation on slopes varying from 0.2 to 5 % Annual rainfall is 1064 mm (Gupta, 1983).

Between 1958 and 1979 the lake lost 60 % of its capacity, despite two partial dredgings (Gupta, 1983). This would correspond with an average inflow of sediment of about 70 m^3/ha/yr. The two streams are connected by an artificial channel, which in itself is a major source of sediment. Concentrations of suspended sediment during the monsoon in the main channel of the Kansal were almost 40 times higher than in three thickly forested headwater catchments (Anonymous, 1977).

Narayana (1987) listed a combination of mechanical, agronomical and forestry measures that have been developed specifically for degraded lands in the Siwaliks by the Central Soil and Water Conservation Research and Training Institute, Dehradun. Application of these measures had a profound effect on the sediment inputs into the lake (S. Chatterji, personal communication).

In this particular case, much of the sediment derived from surface- and channel erosion (Gupta, 1983). As we have seen (Table 12), the former is relatively easy to remedy, whilst the construction of (numerous) spurs, checkdams and debris basins can be very effective in reducing amounts of sediment transported through stream channels (Ram Babu et al. (1980,1981; Narayana, 1987).

Such measures may be expected to be similarly successful in other areas where contributions from surface erosion dominate overall sediment production, such as at the southern plateau (Tejwani, 1985; Narayana, 1987).

Quite a different picture exists, however, for those areas in the Himalaya where riparian mass wasting is the dominant mechanism of sediment supply (Narayan et al., 1983).

Carson (1985) presented a rough calculation of overall (i.e. weighted by land use) surface erosion for the 63-ha Lohore catchment in the Middle Mountains of West Nepal, based on Laban (1979) and Impat (1981)'s comparative data (cf. Table 12). He then assumed an SDR of 50 % and compared the resulting amount of sediment entering the river with a "zonally representative" estimate for stream sediment load of 21 t/ha. Surface erosion in this semi-hypothetical example constituted only 17 % of the total sediment carried by the river (Carson, 1985).

One could argue that in this steep terrain an SDR of 70 % might perhaps have been more appropriate, which would have raised the contribution of surface erosion to 23 %. However, as shown in Table 5, river sediment loads in the Middle Himalayas may be closer to 45 rather than 20 t/ha/yr. Therefore, Carson's estimate of 17 % may well be too high. The inferred rate of sediment production in the riparian zone would amount to some 180 t/ha/yr (Carson, 1985), which is very high in any case.

Although semi-hypothetical, this example is nevertheless instructive, in that it clearly shows that under Himalayan conditions, better management of eroding fields may not necessarily show up as reduced stream sediment loads. If in the present case all surface erosion would have been reduced to that observed under forest, the corresponding reduction in stream sediment would amount to about 10 %.

This of course does not mean that restorative measures should not be applied. They definitely should, but rather in view of the already indicated losses of productivity of cultivated fields in certain parts of the

Himalaya (Shrestha, 1988).

It could be argued that few solid data have been brought forward to illustrate the case. After all, Carson's exercises were largely empirical, although based on considerable field experience. Fortunately, some quantitative data has become available recently, which can be used to test the above view.

Pipal Chaur watershed

The inventory of volumes, types and distribution of slope failures in a 55-ha catchment in the Middle Hills near Dhulikhel, underlain by phyllites and quartzites (Euphrat, 1987), has already been referred to.

Out of 474 landslides with a total volume of 2945 m^3, one major compound slide, located for 90 % in the riparian zone, made up 23 % of the total volume, whilst "naturally occurring" slides (largely in the streamside zone) constituted 80 % of the total volume. Combining volumetric measurements with rough estimates of slide age, Euphrat derived approximate rates of erosion associated with mass wasting per land-use type, which ranged from less than 4 m^3/ha/yr for cultivated land to almost 120 m^3/ha/yr in the riparian zone. Interestingly, the latter value is in the same order of magnitude as that "derived" by Carson (1985) for the Lohore catchment.

In an attempt to assess the importance of mass wasting with respect to contributions by surface erosion, Euphrat (1987) compared his estimates of landslide intensity per land-use category with the surface erosion rates reported by Mulder (1978), Laban (1979) and Impat (1981) for similar types of land use near Pokhara (see Table 12). In addition, as a first approximation of trail erosion, he used an estimate of erosion for a heavily used forest road in the USA as reported by Reid & Dunne (1984). The result of the exercise was that mass movements contributed one-third of the total sediment generated in the area, and various forms of surface erosion the remaining two-thirds.

However, in view of the fact that the area around Pokhara receives more than three (and possibly four) times the amount of precipitation recorded in Euphrat's study area, it is highly unlikely that surface erosion totals for the two areas would be similar.

Also, the erosion rate assumed for trails (1100 t/ha/yr) seems excessive. Schomaker (1988), for example, reported a soil loss from built-up areas (including compacted trails) in Java, Indonesia, under a more aggressive rainfall regime to be in the order of 150 t/ha/yr.

Finally, soil moved by slides in the riparian zone will certainly have more chance of reaching the stream than material eroded from a ridge-crest path. In other words, different weightings would have to be assigned to the various sediment sources in terms of their importance to basin sediment yield.

Therefore, Euphrat (1987)'s conclusion, that stream sediment loads in his area were dominated by surface erosion, is highly questionable. It would be just as easy to "prove", by 1) assigning an SDR of 60 % to all forms of surface erosion as well as to non-riparian slides, 2) by adopting the more modest estimate of trail erosion, and 3) by halving the erosion rates from Pokhara to correct for a much lower rainfall total, that surface erosion in the Pipal Chaur area contributed less than 40 % of the total amount of sediment carried by the stream.

The data on landslide volume and distribution presented by Euphrat are extremely valuable. What is needed to put them in the proper perspective, however, are measurements of surface erosion, that are more representative of regional conditions than the ones derived for the Pokhara area, which is one of the wettest pockets in all of Nepal (Figure 11).

Phewa Tal

A similar analysis can be conducted for the 117 km^2 Phewa Tal catchment in the Middle Mountains near Pokhara, arguably the most researched basin of that size in the entire Himalaya.

Available data include estimates of basin sediment yield based on bathymetric surveys of Phewa lake (Impat, 1981), rates of surface erosion associated with various types of land use (Impat, 1981; to a lesser extent also Mulder (1978) and Balla (1988b)), and observations on mass movement processes (Ramsay, 1985).

Before doing so, however, it must be realized that these data have their limitations. For instance, the results of the bathymetric surveys should be viewed with caution as the difference in storage between the two readings amounted to only 3 % of the total volume of the lake (Impat, 1981) and therefore well within the confidence limits of such an operation. The caveats associated with the erosion plots have been commented upon already in Section IV.2.1 (see also Roels (1985) for a general discussion of the spatial representativity of such installations), whilst Ramsay (1985) himself explicitly stated that his estimates for surface lowering by landsliding should never be used without the prefix "based on a small sample".

Bearing these limitations in mind, the combination of available information presents an interesting picture of sediment production and transport in the Middle Mountain zone of Nepal.

On the basis of the bathymetric surveys Impat (1981) estimated the total amount of sediment trapped in the lake over the period 1976-1979 at 26 m^3/ha/yr, i.e. about 33 t/ha/yr. Assuming a trap efficiency of 90 % this would correspond with an average inflow of sediment into the lake of about 37 t/ha/yr, which is quite comparable to the values quoted for other Mid-Himalayan catchments in Tables 5 and 6 (Section III.6).

Impat (1981) also computed rates of surface erosion for different types of land use by means of the Universal Soil Loss Equation, which were often surprisingly close to actually measured values. Adding up the respective contributions he arrived at a basin-wide estimate of on-site erosion of 89,000 t/yr. Applying an SDR of 0.3 (a reasonable value for a basin this size: Walling, 1983), this would imply that of the 430,000 tonnes (37 t/ha times basin area) of sediment entering the lake each year, about 27,000 tonnes (0.3 times 89,000 t) or 6 % is contributed by surface erosion.

Even if the higher estimate of total surface erosion in the area suggested by Balla (1988b), viz. 135,000 t/yr, is accepted, the contribution of surface erosion remains very modest at less than 10 % of the total sediment yield. The remainder must be supplied by (riparian) mass movements and gully erosion.

Interestingly, on the basis of independent observations of landslide volumes and frequencies in the Phewa valley, Ramsay (1985) arrived at an average estimate of surface lowering by landsliding of about 2.5 mm/yr. In other words, roughly 310,000 m^3 of material is annually transported downhill through various mass wasting processes.According to Ramsay (1985), about 90 % of this material (280,000 m^3) was supplied by a few large failures in groundwater discharge zones. As such, these large features will exhibit a very high transport efficiency and therefore must be responsible for a high proportion of overall sediment movement into the valley bottom river system (Ramsay, 1987a; cf. Brunsden et al., 1981).

Since the central reach of the main river in the area is energy-limited (see Ramsay, 1985 for details), one can only guess as to what fraction of the material thus arriving at the valley floor is transported more or less directly to the lake. Further work is necessary in this regard. Nevertheless, one cannot help noticing the similarity in volumes of sediment generated annually by these large slope failures and those deposited in the lake.

Therefore, the conclusion seems justified that the bulk of the sediment transported by streams in the Middle Himalaya is generated by (a few large) mass movements. Contributions by (accelerated) surface erosion are generally minor.

This finding, of course, has profound implications with respect to the benefits of any upland rehabilitation programmes that can be expected downstream (Hamilton, 1987; cf. III.6).

Fleming (1988) carried out a tentative computation to determine the effect on the siltation rate of the lake of reducing soil loss from overgrazed lands in the Phewa Tal area to a level associated with protected pastures (based on Impat's data).
Not surprisingly in the light of the above considerations, the effect was negligible (about 1 % reduction in siltation).

When Carson et al. (1986) carried out a similar analysis for the Kali Gandaki river basin in West-Central Nepal (11,138 km ^{2}), they arrived at a reduction of 7 % in basin sediment yield following rehabilitation measures.

Apart from the magnitude of downstream effects of upland watershed management activities, there is also a time scale involved (Pearce, 1986). This is clearly illustrated by the fact that the sudden inputs of sediment mobilized during the 1950 earthquake in Assam (Poddar, 1952) remained detectable in the sediment load of the Brahmaputra for more than twenty years (Goswami, 1985; see also Section III.6).

As pointed out by Pearce (1986), there would be very little change for decades in the amounts of sediment carried by major rivers in their lower reaches, even if all man-induced erosion in the uplands could be eliminated at once. The reason for this lies in the fact that there is so much sediment (both from previous man-caused and natural erosion) stored in the system, that this effectively forms a long-term supply.

This idea is also supported by the results obtained from a major land- and stream rehabilitation programme in China, which indicated that under prevailing conditions, reductions in sediment yield upto 30 % could be expected for catchments upto 100,000 km^2 after about two decades (Mou, 1986).

Clearly, the frequently voiced claim that upland reforestation will solve most downstream problems does require some specification of the time scale involved as well.

V CONCLUSIONS

V.1 THE CURRENT SCIENTIFIC CONSENSUS

On the basis of the evidence collated in the preceding chapters, what can we conclude regarding the two questions posed in the introduction, viz. "What is the role of forest and land use in the uplands with respect to flooding, dry-season flows and sedimentation in the plains?", and: "What downstream benefits can be expected in this regard from upland rehabilitation programmes?"

Probably the most important point to be made in this respect, is that one first needs to define the scale for which one's statements are supposed to be valid.

Messerli has proposed a framework, which was summarized by Hamilton (1987) as follows: "at a *local level*, sediment load is strongly influenced by human activity, stream discharge characteristics much less, and, on the whole, human activity has less impact than natural factors on flooding and siltation within the Himalaya. At the *medium level* downstream of the catchment being impacted, we are still uncertain of the quantitative effects of human activity, but the high variability of natural factors dominates both stream discharge and sediment load. At the *macro level* in large basins, human impacts in the upper watershed are insignificant on lowland floods, low flows, and sediment, but these effects can be significantly in-fluenced by human activity in the lower reaches of the river".

Much has been made of the "Himalayan uncertainty" surrounding much of the information on which to base our conclusions. Although this may be true for a number of socio-economic issues, we do feel that the material on bio-physical aspects brought together in the present paper does provide one with a clear and consistent picture, which by and large confirms the contention presented in the last paragraph.

Summarizing: vegetation and land-use practices do exert clear influences on amounts (total water yield) and timing (peakflows, dry-season flows) of streamflow in catchment areas of less than 500 km^2 (cf. Tables 7 and 10). The effect tends to disappear as the area under consideration increases (see below).

Conversion of forest land to agricultural uses (grazing, cropping) will lead to increased total water yields as a result of a reduction in evapotranspirational demand of the new vegetation; dry-season flows may in- or decrease following the conversion, depending on the maintenance of infiltration characteristics of the soil; where the latter deteriorate as a result of poor soil management, more or less severe reductions in low flows can be expected, and vice versa.

Reforesting degraded grass- or croplands with fast-growing trees will generally lead to reduced total and dry-season flows, as the associated increase in water consumption will override the effect of improved rainfall infiltration under Himalayan conditions.

As already indicated, the hydrological effects of land-use manipulations tend to be "diluted" as the area under consideration increases. This holds especially for peakflows, generated by heavy rains. Peak discharges may be increased locally due to poor land management in some parts of a basin, the spatial extent of the phenomenon becomes limited by the size of the rainfall field. Although stormflow volumes generally add up in a downstream direction, the effect is moderated by differences in time lag between tributaries and by the desynchronization imposed by spatial and temporal variations in rainfall.

Truly devastating and widespread flooding is usually the result of an equally large field of extreme rainfall (cf. Figure 33), especially when the event occurs at a time when soils have become wetted up thoroughly by antecedent rains, leaving little opportunity to accommodate the extra water. The process is then governed by ***storage*** capacity rather than ***infiltration*** capacity of the soil. Phrased differently, the presence or absence of a forest cover has become almost negligible under such extreme circumstances.

Floods generated by glacial lake outbursts or the failure of temporary landslide dams are in a class of their own; they are particularly damaging as they carry huge amounts of coarse debris; their occurrence is highly unpredictable and unrelated to land-use practices (cf. Plate 17).

Finally, the extent of flooding in the lowlands is also significantly affected by the occurrence of torrential rains in the plains themselves (cf. Figure 13), the discharging of which from fields is often hampered by high groundwater tables and river stages; "backwater" effects near river junctions (cf. Figures 34-36) or river training works (preventing the lateral spreading of the water) are sometimes important in the generation of high water levels as well.

Although the actual magnitude of major floods in the Ganges-Brahmaputra river basin has most probably not increased significantly over time, the associated economic losses have grown dramatically for a number of reasons. Clearly, the two aspects should be kept separately when discussing the severity of flooding in the area.

Rates of surface ("on-site") erosion are strongly controlled by the degree to which the soil surface is exposed to rainfall or disturbed otherwise; therefore, soil losses from non-terraced cropped fields (e.g in the context of shifting cultivation), and from overgrazed grass- and scrubland (trampling) ranked among the highest recorded in the area (Tables 11 and 12).

Although the physical aspects of soil conservation in the Himalaya are well-understood, the application of certain conservation measures (e.g. closure to grazing, construction of inward-sloping bench terraces, mulching of the surface) may meet with problems of acceptance; given the loss of soil productivity associated with surface erosion on the upper parts of hillslopes, conservation is a must, however; reforestation for erosion combatment should only be promoted if all other measures could be expected to fail.

Rehabilitation of gullied lands can generally only be achieved by a combination of mechanical and vegetative measures; reforestation alone is usually not enough to check this form of erosion.

In many parts of the Himalaya, mass wasting in the riparian zone via lowlevel undercutting by incising streams is the dominant mechanism of sediment supply to the streams; in addition, mass movements on the hillsides are dominated volumetrically by a limited number of very large failures, that are usually found in heavily fractured zones with concentrated groundwater flow; these features are purely geological in nature and are not in any way caused by the presence or absence of a forest cover.

In contrast to these deep-seated slides, the occurrence of shallow (less than 3 metres) landslides is strongly influenced by the presence of a deep-rooting vegetation cover; as such, their number can be expected to decrease in due time following reafforestation of overgrazed or otherwise degraded hillslopes; shallow slips are usually not connected with the overall stream network.

In the light of the above considerations, any effects of land rehabilitation programmes will mainly be felt "on-site" in the form of reduced surface erosion and shallow-landslip incidence; in areas where deep-seated slides dominate sediment production (i.e. in most of the Middle Himalaya), the effects will be negligible or minor at best, even at the scale of relatively small catchment areas (see the computations presented in Section IV.2.3)

Where (accelerated) surface erosion is the most important supplier of sediment to streams (Duns, Siwaliks?, Southern Plateau's), significant improvements have been demonstrated following restoration measures (cf. Table 12).

However, the effect of land improvement schemes on basin sediment yield tends to become more difficult to demonstrate as the size of the catchment area increases. This is due to the fact, that the larger a basin, the more numerous the opportunities to (temporarily) store eroded sediment.

The observation, that the sediment load of the Brahmaputra river in the Assam Valley remained influenced for more than twenty years after the area was hit by a severe earthquake (which released large volumes of sediment into various major tributaries), proves this point rather well.

Clearly, the frequently voiced claim that upland reforestation will solve downstream siltation problems does not pertain to the Himalayan situation. Rather, the bulk of the sediment is generated through natural processes, which are beyond the capacity of man to manipulate. In addition, even if it were possible to cut off all sediment supply to the streams as of today, it would take decades, if not centuries, before any reduction in stream sediment loads would be noticed, especially in the lower reaches of the main rivers.

V.2 GAPS IN OUR KNOWLEDGE

Although a number of general trends have been identified regarding the hydrological and erosion effects of changes in land use in the Himalaya, their detailed quantification at the three scales distinguished earlier requires additional work.

For example, we can be relatively confident that reforestation of degraded scrubland with chir pines or eucalypts, will lead to a reduction in streamflow after several years at the local level. We are much less certain, however, of the amounts of water involved.

Peakflows from small catchment areas in the Dun-Siwalik zone have been reported to respond vigorously to changes in land use. Experimental work of this kind in the Middle Mountains, the area for which this kind of information is most needed, is hardly available. In addition, those studies conducted so far, are largely of a black box type, i.e. not supported by process studies, and therefore difficult to extrapolate.

Overgrazing of grass- and scrublands has been shown to be a major contributor to total accelerated (on-site) erosion, but experiments to determine the carrying capacity of Himalayan grazing grounds seem to be lacking.

At another front, a loss of soil fertility has been noticed in parts of the Middle Mountains. Yet, the rate of the decline in productivity is questionable (J. Dunsmore, personal communication), let alone that a detailed nutrient budget analysis of hill farming systems is available.

These are, we feel, a few of the more pressing questions (from the perspective of the physical sciences) with respect to local environmental management problems in the Himalaya, . By their very nature, these questions will need to be addressed at the "micro" scale, but within the various major physiographic zones, so as to ensure their proper positioning within the regional framework.

Naturally, most of this work will need to be of an interdisciplinary nature, and should take full advantage of locally available environmental knowledge. It is suggested, that such research be conducted in a number of carefully selected, well-instrumented catchment areas, which would enable one to use statistically sound techniques for the evaluation of the hydrological effects of changes in land use or management (Figure 44).

In order to separate land-use effects from climatic effects, such studies need to be conducted over a fairly long period of time. However, much can be learned from well-designed short-term process studies (Gilmour et al., 1987), especially when carried out in the context of a catchment experiment (Sklash et al., 1986; Bruijnzeel, 1987).

As we have seen in the preceding chapters, an evaluation of the role of land use in determining streamflow and sedimentation patterns at the meso- and macro-scales, is much more difficult. Not only will there be a variety of land-use types in a single basin, each exerting a specific influence, but also, and arguably more importantly, there will be major variations in the spatial and temporal distribution of rainfall. As such, the issue can hardly be approached experimentally.

Ives et al. (1987) formulated an outline for a research strategy, which aimed at determining the downstream impacts of "perceived" human activity in the mountains. They proposed the selection of several major Himalayan basins for the long-term monitoring of streamflow and sediment patterns, to be linked with a "minimal network of observation stations" in the various physiographic zones. This should in turn be supplemented by a number of "experimental watersheds at different altitudes, and with different slope angles and aspect, vegetation and land-use types", where "routine climatological, hydrological, and geomorphological data should be collected from sites selected to become focal points for a full range of human science research".

However, at the meso- to macro-scales, the "routine collection of basic hydrological and climatic data" has been in operation for decades in the Indian Himalaya (see various reports summarized in Anonymous, 1981a; Sakthivadivel & Raghupathy, 1978; Seth & Datta, 1982; Dhar et al., 1982b). As such, if the problem of highland-lowland interaction is to be approached through time-series analysis of hydrological data, it is obviously advantageous to use data that go back in time as far as possible. Also, it will be much more difficult to distill any (further!) deterioration in river regimes from a recently initiated monitoring programme (cf. Figure 43), as shown by similar attempts in Thailand (Dyhr-Nielsen, 1986) and Southern China (Qian, 1983).

As the sole possessors of such long-term streamflow records in the region, the Indian political and scientific communities have a special responsibility in this respect.

More promising may be the organization of a network of representative (erroneously called "experimental" by Ives et al., 1987; see Simmers, 1984) catchments in the Himalaya. However, such a strategy will only yield meaningful data (i.e. in the context of land-use influences), if it is lifted from the level of routine collection of basic data (as proposed by Ives et al., 1987) to truly experimental work involving actual "treatments" (see Chapter IV.1).

Although it is beyond the scope of the present report to discuss these matters in any detail, it would seem that ICIMOD could play a very useful coordinating role in this respect (Alford, 1988b).

VI. REFERENCES

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ERRATA ICIMOD OCCASIONAL PAPER NO. 11

Page VIII, Table 9: "log $K_s$at" should read: "log K_{sat}"

Page 7, Figure 3, "reserved" should read "reversed"

Page 12, Plate 5 has been printed upside down

Page 23, Caption Figure 9a, add: "all values in cm/yr"

Page 30 Figure 13 (Maximum 24-hour rainfall totals): the values read as follows (from NW to SE):

Chakrata (192); Dehradun (152); Hardwar (495); Roorkee (272); Delhi (266); Agra (263); Bhopal (233); Pauri (173); Lansdowne (258); Nagina (823); Dhampur (497); Moradabad (161); Almora (221); Nainital (185); Silgarhi (156); Dhangadi (283); Lucknow (51); Jomosom (87); Pokhara (278); Butwal (403); Varanasi (350); Kathmandu (386); Patna (366); Ranchi (231); Taplejung (95); Dhankuta (265); Bharakshetra (405); Darjeeling (640); Calcutta (369); Gauhati (233); Shillong (415); Cherrapunji (1036); Dhaka (326); Chittagong (417); Dibrugarh (745)

Page 66, Figure 36, "S A N" should read "S O N"

Page 80 Figure 43c represents variables of Figure 43a (not b)

Page 95 Figure 53b. Discharge in thousands of m^3/sec (not m^3/sec)